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DIFFERENTIATION OF THE DOLERITES OF TASMANIA. I

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ABSTRACT

The dolerites of Tasmania are close kin to those of the Karroo and especially to those of Antarctica. They occur as sills, sometimes more than 1,000 feet thick, and as great dikelike intrusions a mile or more wide in the horizontally bedded Permo-Carboniferous and Trias-Jura sediments of the island. Their outcrops cover an area of more than 6,000 square miles and may previously have extended over twice that area. The sills, in particular, with their pronounced columnar structure, occupy a prominent place in the Tasmanian landscape.

At the time of its intrusion the dolerite magma was completely liquid, and analyses of chilled margins from widely separated localities show that it was homogeneous in composition. It was a saturated or tholeiitic basalt magma but differed from typical tholeiites by being richer in SiO_2, Al_2O_3, and CaO and poorer in FeO, TiO_2, Na_2O, K_2O, and P_2O_5. The only perfect match is provided by the Antarctic dolerites.

The dolerites underwent fractional crystallization and gravitative differentiation after emplacement, which led to a progressive enrichment of the later-formed pyroxenes in iron as differentiation progressed. Two immiscible series of pyroxenes crystallized side by side throughout—at first, magnesia-rich augite and orthopyroxenes, later iron-enriched augites and stable iron-magnesia pigeonites ("plutonic" pigeonites). The enrichment of the pigeonites in iron progressed more rapidly than the enrichment of the

INTRODUCTION

The sills and dikes of Mesozoic dolerite in Tasmania are closely akin, both in age and in composition, to the great dolerite sills of Antarctica and the Karroo System of South Africa. Their outcrops cover approximately 6,200 square miles out of the total 26,000 square miles of the island (Fig. 1); and from the distribution of dolerite outliers it is possible that they previously covered an area of more

augites in iron. This behavior and the immiscibility of the two series of pyroxenes can be explained by reference to their atomic structure and the much greater ease with which a small Fe^{++} ion can replace an Mg^{++} ion of similar ionic radius as compared with a much larger Ca^{++} ion. It follows from such an argument that the immiscibility gap found between magnesia-rich augites and magnesia-rich orthopyroxenes and clinoenstatites should extend between the iron-rich varieties as well. Supporting evidence that this is so is found in the behavior of the pyroxenes of the Mount Wellington sill, Tasmania, of the Skaergaard intrusive in East Greenland, of the extrusive rocks of the Huzi Volcanic Zone, Japan, and of the Palisade sill, New Jersey. From these data a new triangular diagram expressive of the courses of crystallization of pyroxenes is drawn.

Identical fractional crystallization and gravitative differentiation have given a contrast in trends of differentiation in the dolerite sills compared with the dikelike bodies. This contrast arises from the different forms of the chambers in which differentiation occurred. In the closed chambers of the sills, sinking of early-formed magnesia-rich pyroxene caused a slight absolute enrichment in iron of the residual magma displaced into the upper parts of the sills. The iron-rich pyroxene that formed could not sink past the layer of accumulated early-formed minerals and so was retained in the upper parts of the sills to form peculiar iron-rich rocks. In the dikelike intrusions, which were subjacent bodies (in the sense defined by Daly), there was no layer of accumulated early-formed crystals to prevent the sinking of the later-formed iron-rich pyroxenes. As a result, the concentration of iron in the residual magma was dissipated, and the trend of differentiation in the dikelike masses gave rise to andesitic rocks that resemble in many respects the two-pyroxene andesites of Japan.

This dependence of the trend of differentiation on the form of the differentiation chamber is of some significance in petrogenesis and reconciles apparently conflicting views like those held by C. N. Fenner and N. L. Bowen as to the ultimate effects of the enrichment of the residual liquids of basalt in iron—a trend which appears to characterize all basaltic magmas. Fenner's conclusion that this trend should give rise to absolute enrichment of the residual liquid in iron holds true for closed chambers of differentiation, of which the Skaergaard intrusion in East Greenland is the example par excellence. The same trend is found on a lesser scale in the dolerite sills of Tasmania and the Palisade sill of New Jersey. Bowen's conclusion that the enrichment of the residual liquids of basalt in iron leads merely to an enrichment of the later-formed ferromagnesians in iron relative to magnesium but not to an absolute enrichment in iron holds true for basaltic magmas that differentiated in subjacent chambers. Thus an olivine-basalt magma gives rise to "ferro-gabbro" in a closed chamber, while a tholeiitic magma gives rise to peculiar iron-rich rocks in a closed chamber and to normal calc-alkaline rocks in a subjacent chamber. The relative volume of calc-alkaline rocks produced in this way is, however, no greater than the proportion of trachytic rocks produced from olivine-basalt, so that it seems unlikely that the great bulk of calc-alkaline rocks are derived simply by the differentiation of tholeiitic magma.

than 15,000 square miles. The sills were often more than 1,000 feet thick, and some of the partially eroded sills, such as that capping Mount Wellington, still exceed this thickness. Assuming an average original thickness of 1,050 feet (i.e., one-fifth of a mile) the volume of dolerite obtained for the present outcrops is about 1,200 cubic miles and for its possible previous extent about 3,000 cubic miles.

The resistant nature of the dolerites, combined with their sill-like form, makes them prominent features of the Tasmanian landscape, especially in regions like the Central Plateau and the Ben Lomond Plateau, which have been elevated by later block faulting and now

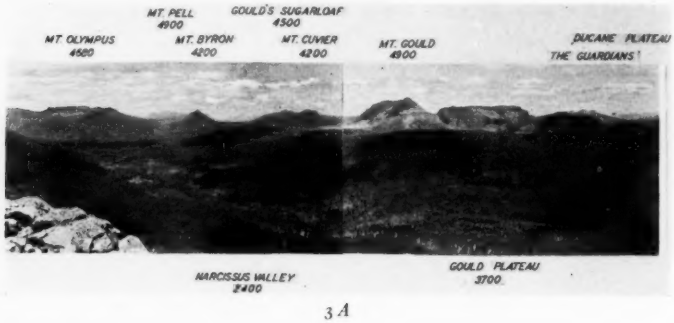
stand at from 2,500 to 5,000 feet above sea-level. In both of these areas erosion has exposed the dolerite and left it as a flat-topped capping to the plateaus. The columnar structure of the sills has

FIG. 1.—Locality sketch map, showing the approximate distribution of outcrops of Mesozoic dolerite in Tasmania (taken from the geological map of Tasmania printed in *Bull. Geol. Soc. Tasmania*, Vol. XLIV [1938]).

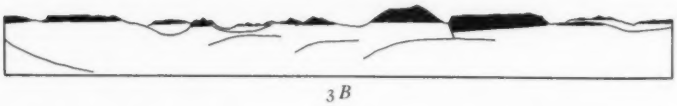
given rise to steep or precipitous cliffs at the edges of the plateaus, where the columns have collapsed into coarse scree deposits (Fig. 2).

On the west side of the Central Plateau, between Lake St. Clair

FIG. 2.—"The lofty, craggy and precipitous battlements of Ben Lomond." The eastern edge of the great sill capping the Ben Lomond Plateau. (Photo by H. J. King.)



3 *A*



3 *B*

FIG. 3.—(*A*) View of the dissected dolerite sill in the Lake St. Clair district, looking southeast down the Narcissus Valley toward the lake from Traveller's Range. In the middle distance is Gould Plateau, with a shadow on it, which marks the top of the horizontally bedded sediments that underlie the sill. (Photo by R. H. Major.) (B) Shows the remnants of the sill (black) diagrammatically.

and Cradle Mountain, erosion has been severe on account of the higher rainfall and the Pleistocene glaciation, which affected this region, and the dolerite sills have been dissected to varying degree. In the early stages of dissection the sills give rise to groups of mountains with characteristic columnar margins, all rising to about the same level above a partially dissected plain of more or less horizontally bedded sedimentary rocks (Fig. 3). At a more advanced stage

FIG. 4.—Mount Wellington (4,150 ft.), near Hobart, is the isolated portion of an eroded sill. It caps horizontal Trias-Jura sediments and is still more than 1,000 feet thick. (Photo by A. N. Lewis.)

the mountains become isolated, and the columnar dolerite remains as a capping above a deeply eroded sedimentary base. With further erosion the area of the dolerite cap is reduced more and more (Fig. 4), until only small isolated residuals remain, each perched on a "pedestal" of sedimentary rocks, well above the general level of the country (Fig. 5). In places, as at the northern end of Lake St. Clair, erosion has gone as far as to remove the dolerite completely but has not yet worn down the empty pedestal. Dolerite outliers are found all along the present margins of the dolerite of the Central Plateau and particularly in the southeastern part of the island.

Where the dolerite sills are exposed to marine erosion, as in the drowned areas of southeastern Tasmania, they form magnificently rugged coasts.

In the lower area of the Midlands the dolerites are still prominent in the landscape, but they usually take the form of abrupt, steep-sided ridges (Fig. 6), which lack the columnar structure of the

FIG. 5.—Cradle Mountain (5,069 ft.), northwest Tasmania, is a remnant of a once extensive dolerite sill overlying nearly horizontal Permo-Carboniferous beds, which lie upon the peneplaned surface of the pre-Cambrian. Lake Dove is in the foreground. This represents an advanced stage in the dissection of a dolerite sill. (Photo by H. J. King.)

dolerites on the plateaus. This is because in the Midlands the dolerites occur in the form of great dikes, with more or less vertical walls and without the vertical columns of the sills, since cooling proceeded from the sides rather than from the base and top.

INTRUSION OF THE DOLERITES

The rocks into which the dolerite magma was intruded consist of a pre-Cambrian and Lower Paleozoic basement, overlain by sediments of Permo-Carboniferous and Trias-Jura age. The basement

rocks are folded sediments and metamorphics which have been intruded by igneous rocks—chiefly granites and acid porphyries. They were involved in a presumably Devonian orogeny, which was followed by a period of peneplanation. The younger sediments were deposited on the surface of the peneplain and aggregate several thousands of feet in thickness in some regions, particularly in the southeast. These younger sediments have been tilted at angles up to

FIG. 6.—Barn Bluff (5,115 ft.) in northwest Tasmania, seen from Cradle Mountain, consists of a small dolerite residual perched on a "pedestal" of sedimentary rocks. This is the penultimate stage in the erosion of a sill. (Photo by Spurling.)

10° by faulting and possibly by the intrusion of dolerite sills along their bedding planes, but they have not been involved in folding movements.

There is a distinct relationship between the distribution and mode of occurrence of the dolerite and the structure of the intruded sediments.[1] Where the dolerite intruded the folded basement rocks, the intrusions took the form of dikes, generally of small volume. On the other hand, where it intruded the younger unfolded sediments, it

[1] A. McIntosh Reid and Q. J. Henderson, "The Avoca Mineral District," *Bull. Geol. Surv. Tasmania*, No. 40 (1929), p. 17.

formed chiefly sills, often of great volume, occasional laccoliths, and, especially in the Midlands district, large ramifying dikelike bodies. The sills sometimes developed along the unconformity at the base of the unfolded sediments, as at Mount Anne[2] in southwest Tasmania and at Mount Sedgwick[3] in western Tasmania.

This relationship indicates that the dolerite magma rose rapidly

TABLE 1

	1	2	3	4	5	6
SiO_2	52.50	52.60	52.54	53.00	52.75	52.50
Al_2O_3	16.38	16.80	16.89	17.12	15.89	14.30
Fe_2O_3	0.77	0.08	0.27	1.02	0.81	0.15
FeO	8.12	8.64	8.09	7.27	8.50	8.63
MgO	6.77	6.76	6.36	6.25	6.54	7.15
CaO	11.20	11.60	11.10	11.30	12.30	10.55
Na_2O	1.54	1.34	1.82	1.30	1.64	1.82
K_2O	0.80	0.80	0.83	0.78	1.04	1.15
H_2O+	0.47	0.18	0.21	0.38	0.18	1.46
H_2O-	0.40	0.31	0.76	0.58	0.20	1.25
CO_2	nil	nil	nil	nil	nil	nil
TiO_2	0.65	0.48	0.65	0.62	0.51	0.60
P_2O_5	0.05	tr.	tr.	tr.	tr.	0.02
MnO	0.12	0.28	0.18	0.16	0.08	0.11
Total	99.77	99.87	99.70	99.78	100.44	99.69

1. Chilled base, actual contact on Pinnacles Road, Mount Wellington, 3,125 feet above sea-level.
2. Chilled top, 650-foot level on Proctor's Road, Mount Nelson.
3. Chilled base, 20–30 feet above contact, Mount Sedgwick, east side.
4. Chilled base, about 30 feet above contact, Ben Lomond, above Storey's Creek township.
5. Chilled base, contact at 475-foot level on south side of Gunning's Sugarloaf.
6. Chilled base, at eastern end of Mount Nicholas Range.
A. B. Edwards, analyst.

into the upper parts of the crust through fissures and planes of weakness. Further evidence of this is provided by the remarkably uniform composition of the chilled margins of the sills and the dikelike bodies (Table 1). The mineralogical composition of the chilled phases indicates that at the time of intrusion the magma was at a temperature of not less than 1,000° C., but the amount of contact metamorphism induced by it was extremely slight. In part this was

[2] A. N. Lewis, "Notes on a Geological Reconnaissance of Mount Anne and the Weld River Valley, South-West Tasmania," *Papers Roy. Soc. Tasmania* (1923), pp. 24–25.

[3] A. B. Edwards, "On a Remnant of a Stripped Peneplain of Palaeozoic Age at Mount Sedgwick in Western Tasmania," *Papers Roy. Soc. Tasmania* (1941).

because many of the intruded sediments were composed largely of minerals stable at temperatures of 1,000° C.; but, even where the invaded rock was limestone, shale, or coal, the amount of metamorphism was slight,[4] showing that the heat supplies of the intrusions were limited.

Sills of dolerite were intruded at all horizons from the base of the Permo-Carboniferous to the top of the Trias-Jura,[5] wherever a feeder cut a horizontal plane of weakness. Originally, therefore, the sills were distributed over a maximum vertical range of at least 8,000 feet. For the dolerite magma to have risen rapidly over such a range of height and thrust itself along the bedding planes of the intruded sediments, it must have been under great pressure, but it seems rarely, if ever, to have reached the surface. Tuffs are recorded in the Triassic sediments near Hobart,[6] but the volume of magma, if any, which may have been extruded was not at all comparable with the amount that failed to reach the surface, or there would be some indication of its former existence. This means that the fissures up which the dolerite rose developed from below upward but failed to break through to the surface. The subcrustal regions must have undergone expansion or have been in a state of tension to cause such fissuring in the crust. Near the surface, where the sediments were but newly formed, the tension was relieved, not by breaking, but by stretching, and presumably by a gliding of the horizontal beds over one another. The dolerite magma, under great pressure as a result of this state of internal expansion, flowed up the fissures as they opened and thrust itself into the "gliding planes" in the sediments, where, no doubt, it acted as a lubricant. In other places, where the fissure ended "blind," the dolerite magma filled it and so gave rise to dikes such as occur in the Midlands.

[4] T. Stephens, "On an Intrusion of Diabase into Permo-Carboniferous Rocks in Frederick Henry Bay," *Quart. Jour. Geol. Soc. London*, Vol. LVI (1900), pp. 333–36; P. B. Nye, *Underground Water Supply Papers, Geol. Surv. Tasmania*, No. 1 (1921), pp. 52–54; No. 2 (1922), p. 38; No. 3 (1924), p. 35; "The Coal Resources of Tasmania," *Mineral Resources No. 7, Geol. Surv. Tasmania* (1922), pp. 35–36, 82, 139, 173, 177, 210.

[5] Lewis, "The Isostatic Background of Tasmanian Physiography," *Papers Roy. Soc. Tasmania* (1927), pp. 1–24.

[6] A. N. Lewis and A. H. Voisey, "A Record of Volcanic Activity in Tasmania during Triassic Times," *Papers Roy. Soc. Tasmania* (1938), pp. 31–39.

PETROLOGY

Previous Work

The first petrographical examination of the Tasmanian dolerites was made by W. H. Twelvetrees[7] in 1896 on the dolerite of Cataract Gorge, Launceston. More comprehensive studies in 1898[8] and 1899[9] of specimens from many parts of Tasmania revealed their remarkable similarity and led to the conclusion that "the rock all over the Colony belongs to one and the same geological period." The essential minerals were described as augite and basic plagioclase, while olivine, apatite, ilmenite, magnetite, pyrite, biotite, quartz, and orthoclase were found as occasional accessory minerals. The specimens examined were all holocrystalline and varied in texture from ophitic to intergranular, the groundmass being essentially feldspathic. Later Twelvetrees[10] recorded the occasional presence of hypersthene and enstatite in these rocks, together with granophyric intergrowths of quartz and orthoclase in dolerites in the vicinity of Hobart. He also noted two variations from the otherwise "monotonously uniform" structure of the rock throughout the island, namely, a coarsening toward a gabbroid appearance and an occasional tendency to be porphyritic near the margins.

In 1907 Osann,[11] in an annex to W. Wahl's[12] investigation of the enstatite-augites, described the occurrence of an almost uniaxial variety of enstatite-augite in a specimen of dolerite from Cataract Gorge, Launceston. His attempts to separate this mineral from the other pyroxenes in the rock were not successful, but an analysis of the composite pyroxenes (Table 10) confirmed his optical determina-

[7] "Microscopical Studies of Tasmanian Rocks," *Papers Roy. Soc. Tasmania* (1896), pp. 89–91.

[8] W. H. Twelvetrees and W. F. Petterd, "The Igneous Rocks of Tasmania," *Trans. Australian Inst. Min. Eng.*, Vol. V (1898), pp. 98–116.

[9] Twelvetrees and Petterd, "On Mesozoic Dolerite and Diabase in Tasmania," *Papers Roy. Soc. Tasmania* (1899), pp. 47–55.

[10] "On the Nomenclature and Classification of Igneous Rocks in Tasmania," *Rept. Australian Assoc. Adv. Sci.*, Vol. IX (1902), pp. 287–92.

[11] "Über einen Enstatit-Augite führenden Diabas von Tasmanien," *Centralbl. f. Min.* (1907), pp. 705–11; also, translation by W. H. Twelvetrees in *Ann. Rept. Dept. Mines Tasmania, 1907*, pp. 107–11.

[12] "Die Enstatit-Augite," *Min. u. petrog. Mitt.*, Vol. XXVI (1906), pp. 1–131.

tion of the mineral. A chemical analysis of the Cataract Gorge dolerite was included in the paper (Table 2, No. 1).

In the previous year F. P. Paul[13] had analyzed a specimen of dolerite from the North-West Bay (Table 3, No. 2), which also contained pyroxenes of different optic axial angle. The presence of quartz in the groundmass of this rock led Paul to classify it as a "konga-diabase." Twelvetrees,[14] commenting on this rock, states

TABLE 2

	1	2
SiO_2	52.49	56.74
Al_2O_3	16.44	15.46
Fe_2O_3	2.60	3.08
FeO	5.30	7.58
MgO	6.18	2.54
CaO	11.71	7.64
Na_2O	2.06	3.08
K_2O	1.09	1.59
H_2O+	0.15	1.28
H_2O-	1.42	
TiO_2	0.62	1.26
P_2O_5	tr.	0.15
MnO	tr.	tr.
Total	100.06	100.40

1. Dolerite, or enstatite-augite–bearing diabase, from Cataract Gorge, Launceston. Dittrich, analyst.

2. Quartz-dolerite, or konga-diabase, from North-West Bay. F. P. Paul, analyst.

that "the quartzose variety, though frequently met with in the south of the island, is not quite typical for Tasmanian diabase." The present study has demonstrated that Paul's "konga-diabase" is a product of the differentiation *in situ* of the dolerite magma, which approximated in composition to the Launceston specimen described by Osann.

Osann's determination of the occurrence of enstatite-augite in the

[13] "Beiträge zur petrographischen Kenntniss einiger foyaitischentheralitischen Gesteine aus Tasmanien," *Min. u. petrog. Mitt.*, Vol. XXV (1906), p. 267.

[14] "Outline of the Geology of Tasmania," *Rept. Sec. for Mines for 1908, Tasmania*, p. 154.

dolerite was confirmed by J. A. Thomson[15] and later by W. N. Benson,[16] who discovered that there were three varieties of pyroxene in the dolerite of Cradle Mountain. In a specimen from the lower part of the dolerite capping of Cradle Mountain, Benson observed enstatite and augite, while in a specimen from the summit of the mountain he found an almost uniaxial enstatite-augite dominating the other two. He also noted that the dolerite was chilled at its contact with the underlying sediments. Similar chilled phases were remarked by Nye[17] wherever the dolerites of the Midlands district are in contact with intruded sediments. Nye noted that the dolerites became increasingly coarse in texture as one progressed from the chilled margins toward the centers of the masses. The widespread occurrence of a chilled margin in the dolerites in other parts of Tasmania is confirmed by Lewis.[18]

The Present Work

The variations noted in these several studies indicate that the Tasmanian dolerites are not so uniform as their appearance in the field leads one to expect. This study was undertaken, therefore, to find out to what degree they showed variation and whether such variation was due to differentiation *in situ* or before intrusion or both. The great columnar cliffs, so characteristic of the edges of these dolerite sills (Fig. 2), offer excellent exposures for testing these possibilities. A number of suitable localities widely spaced throughout Tasmania were selected for examination. These were visited either by myself or by friends, and representative collections were made at intervals from top to bottom of the exposures. Chemical and petrographical examination of these collections have revealed two striking facts: (1) the dolerite magma, at the time of its intru-

[15] "On Rock Specimens from Central and Western Australia, Collected by the Elder Scientific Exploring Expedition of 1891–92," *Jour. and Proc. Roy. Soc. New South Wales*, Vol. XLV (1911), p. 306.

[16] "Notes on the Geology of the Cradle Mountain District," *Papers Roy. Soc. Tasmania* (1916), p. 33.

[17] *Underground Water Supply Papers*, *op. cit.*, Nos. 1, 2.

[18] "Notes on a Geological . . . ," *op. cit.*; "The Isostatic Background . . . ," *op. cit.*

sion, was completely liquid and of an extremely uniform composition and (2) it underwent differentiation *in situ* in a remarkably uniform manner.

The Dolerite Magma

The dolerite sills and dikes show chilled margins wherever they are in normal (that is, not faulted) contact with the intruded sediments. Chilled margins occur at both contacts of the large dikes in the Midlands; and the bottoms of sills invariably consist of a chilled base, 30 or more feet thick, often masked by scree deposits. The chilled tops of the sills, on the other hand, are not often preserved. Lewis[19] has described such a chilled top at Mount Anne, in southwest Tasmania, where a dolerite sill 1,200–1,500 feet thick has intruded an area of quartzites. The chilled top of the sill, which Lewis estimated to be about 50 feet thick, has been protected from erosion by a cover of quartzite. A small remnant of a chilled top, 20–30 feet thick, with portion of its sedimentary cover, is preserved on a downfaulted block at Mount Nelson, near Hobart, and is exposed where the 650-foot contour crosses Proctor's Road up the mountain. Another remnant of a chilled top is reported to occur on the Ben Lomond Plateau.[20]

These chilled margins provide the nearest approach to the nature and composition of the magma from which the intrusions were derived. Specimens of such chilled margins were collected from a number of widely separated localities. All those examined are fine-grained, holocrystalline rocks, with an intergranular texture, and consist essentially of basic plagioclase and pyroxene. The plagioclase (An_{70}) occurs as stumpy laths, about 0.10×0.25 mm., and the pyroxene occurs as stumpy prisms and irregular grains of augite and bronzite of about the same dimensions. This uniformity of composition is reflected in their chemical analyses (Table 1). The grain size varies slightly from specimen to specimen and is finest in those taken from the immediate vicinity of a contact.

In the specimen from the absolute contact of the Mount Wellington sill, the bulk of the pyroxene is so fine grained that its optical

[19] "Notes on a Geological . . . . ," *op. cit.*

[20] Q. J. Henderson, personal communication.

properties could not be determined, but an analysis of the composite pyroxene from this specimen (Table 10, No. 1) indicates that it is a pigeonite such as is found in the groundmass of some basalts.

The bronzite was the first mineral to form and presumably began to crystallize prior to, or during, the intrusion of the magma, because in each rock there are occasional microphenocrysts of bronzite, generally fringed with a narrow rim of granular augite. In some sections the plagioclase shows the same tendency, but not to the same degree. Occasionally, as in the Mount Nicholas sill in northeastern Tasmania, crystallization was slightly more advanced prior to intrusion, because specimens from that locality contain considerably larger and more numerous phenocrysts of enstatite or bronzite. This explains the higher MgO content of this rock (Table 1, No. 6).

The Mount Sedgwick rock is also unusual. Though not apparent from its analysis (Table 1, No. 3), small, much-corroded crystals of olivine, up to 0.5 mm. in diameter, are scattered sparsely through the specimen, sometimes enveloped in narrow rims of pyroxene. At higher levels in the Mount Sedgwick laccolith these olivine crystals have been almost completely resorbed.

It is apparent, therefore, that the dolerite magma was almost completely liquid at the time of its intrusion and was on the point of crystallization. As the chemical analyses show, it was a tholeiitic, or saturated magma, of a very uniform but slightly unusual composition; and it tended to give rise to saturated minerals from the outset. In Table 3 the average composition of the undifferentiated Tasmanian dolerite, as given by the analyses of Table 1, is compared with the chemical composition of dolerites of more or less similar age in other parts of the world. The resemblance to the dolerites of South Victoria Land,[21] King George Land,[22] and Adelie Land[22]—those parts of the Antarctic continent immediately facing Tasmania—is most striking, the more so since other analyses of the Antarctic dolerites

[21] G. T. Prior, "Report on the Rock Specimens Collected during the 'Discovery' Antarctic Expedition, 1901–4," *Nat. Antarctic Exped., 1901–4, Natural History*, Vol. I, *Geology*, p. 137; Benson, "Report on the Petrology of the Dolerites Collected by the British Antarctic Expedition of 1907–8," *Repts. British Antarct. Exped., 1907–9, Geology*, Vol. II, pp. 153–60.

[22] W. R. Browne, "The Dolerites of King George Land and Adelie Land," *Repts. Australian Antarct. Exped., 1911–1914, Ser. A.*, Vol. III, *Geology*, Part 2 (1923).

closely resemble certain of the differentiated phases of the Tasmanian dolerites. The Tasmanian and the Antarctic dolerites are characterized chemically by the following distinctive features: (1) relative richness in CaO and Al_2O_3, and paucity in Na_2O; (2) rela-

TABLE 3

	1	2	3	4	5	6	7	8
SiO_2	52.65	53.05	54.16	53.16	51.91	52.25	53.21	50.25
Al_2O_3	16.23	16.90	15.08	15.01	15.31	14.60	13.95	13.76
Fe_2O_3	0.51	0.79	0.79	1.27	0.98	0.84	1.24	3.87
FeO	8.21	6.69	8.08	8.29	9.31	9.89	8.93	8.50
MgO	6.64	6.91	7.14	7.45	7.52	6.95	7.19	5.42
CaO	11.34	11.56	10.57	10.36	9.71	9.71	8.95	9.09
Na_2O	1.58	2.05	1.60	2.22	2.30	2.21	2.65	2.42
K_2O	0.90	0.97	1.11	1.07	0.79	0.96	1.13	0.96
H_2O+	0.48	0.49	0.36	0.56	0.93	0.71	0.49	1.51
H_2O-	0.85	0.43	0.20		0.15	0.32	0.22	0.76
CO_2	nil	.....	tr.	.....	.....	.....	.....	.....
TiO_2	0.58	0.65	0.70	0.16	1.25	1.10	1.73	2.39
P_2O_5	0.01	tr.	tr.	0.08	0.18	0.22	0.21	0.26
MnO	0.15	0.07	0.14	0.15	0.08	0.45	0.16	0.16
Total	100.13	100.61	99.93	99.87	100.42	100.21	100.13	100.31

1. Average undifferentiated Tasmanian dolerite (six analyses).

2. Dolerite from Horn Bluff, King George Land, Antarctica (W. R. Browne, *Science Repts. Australian Antarct. Exped., 1911–14, Ser. A.*, Vol. III, *Geology*, Part 3 [1923]).

3. Aphanitic quartz-dolerite, erratic No. P. 336, Cape Royds, South Victoria Land, Antarctica (W. N. Benson, *Repts. British Antarct. Exped., 1907–9*, Vol. II, *Geology*, p. 157).

4. Quartz-dolerite from the Cuyuni River, British Guiana (J. B. Harrison, *Geology of the Goldfields of British Guiana*, p. 23).

5. Average undifferentiated Palisade diabase (F. Walker, *Bull. Geol. Soc. Amer.*, Vol. LI [1940], p. 1081).

6. Average Karroo dolerite (R. A. Daly and T. F. W. Barth, *Geol. Mag.*, Vol. LXXI, [1930], p. 101).

7. Dolerite of Downes Mountain, Calvinia, South Africa (F. Walker and A. Poldervaart, *Trans. Geol. Soc. South Africa*, Vol. XLIII [1941], p. 159).

8. Average Whin sill magma (A. Holmes and H. F. Harwood, *Min. Mag.*, Vol. XXI [1928], p. 539).

tively low iron content; (3) relatively low TiO_2 content; (4) practical absence of P_2O_5; (5) low MnO content. These characteristics differentiate them from other well-known dolerite formations, like those of the Palisade sill in New Jersey,[23] the Karroo System in

[23] F. Walker, "Differentiation of the Palisade Diabase, New Jersey," *Bull. Geol. Soc. Amer.*, Vol. LI (1941), p. 1081.

South Africa,[24] and the Whin sill in England.[25] These form a second group, which is (1) richer in Na_2O and poorer in CaO and Al_2O_3, (2) richer in iron, (3) richer in TiO_2, (4) richer in P_2O_5, than the first group. The dolerite from the Cuyuni River, in British Guiana (Table 3, No. 4) is intermediate between the two but resembles the Tasmanian and Antarctic dolerites more closely than the others.

Differentiation of the Dolerite Magma

Because of the peculiar mineralogical composition of the dolerite sills and dikes, their differentiation has been studied largely by chemical analyses. The dolerite consists essentially of plagioclase and pyroxenes, and the chief mineralogical change that accompanies differentiation is a progressive enrichment of the pyroxenes in iron relative to magnesium. The compositions of the various pyroxenes present are such that the variations in their optical properties are not readily detected in thin sections. Their optic axial angles, as far as can be observed without a universal stage, alter but slightly over the whole range of their change of composition; and, although there is an increase in refractive index and a change in extinction angle, these properties are not easily assessed in thin sections or even in the crushed rock, because two or more varieties of pyroxene are often intergrown with one another. Examination of thin sections, unaccompanied by chemical analyses, gives the impression, therefore, that, apart from an increase in grain size in the more slowly cooled portions of the intrusions, the mineral composition of the dolerites is "monotonously uniform."[26] Sufficient chemical analyses show that this is not so. Moreover, the chemical variations so revealed can be interpreted readily in terms of the few essential minerals composing the dolerites.

I. THE DOLERITE SILLS

1. MOUNT WELLINGTON

The 1,000-foot-thick sill capping Mount Wellington, 10 miles west of Hobart (Fig. 4), was selected as a typical sill. It lies more or

[24] R. A. Daly and T. F. W. Barth, "Dolerites Associated with the Karroo System, South Africa," *Geol. Mag.*, Vol. LXVII (1930), p. 101.

[25] A. Holmes and H. F. Harwood, "The Age and Composition of the Whin Sill and the Related Dykes of the North of England," *Min. Mag.*, Vol. XXI (1928), p. 539.

[26] Twelvetrees and Petterd, "On Mesozoic Dolerite . . . . ," *op. cit.*

less horizontally upon Triassic beds of the Ross sandstone stage, and the cuttings of the motor road to the summit provide an almost continuous exposure of fresh rock from top to bottom of the sill. The

TABLE 4

ANALYSES OF MOUNT WELLINGTON DOLERITE SILL

	1	2	3	4	5	6	7	8	9	10
SiO_2	52.50	52.95	52.50	52.60	52.25	53.30	54.90	54.70	55.50	57.05
Al_2O_3	16.38	15.29	13.39	14.99	17.90	17.40	17.41	17.48	17.63	16.95
Fe_2O_3	0.77	1.37	0.90	0.85	0.12	1.26	0.56	0.25	0.05	2.60
FeO	8.12	7.70	8.72	8.78	8.64	8.50	8.50	8.91	9.94	8.64
MgO	6.77	7.49	9.16	8.00	5.33	4.38	3.33	2.86	2.35	1.77
CaO	11.20	11.60	12.25	11.95	11.51	11.75	9.75	9.70	8.70	8.60
Na_2O	1.54	1.10	0.80	1.30	2.07	1.39	2.17	2.33	2.17	2.09
K_2O	0.80	0.72	0.70	0.80	1.04	0.82	1.15	1.33	1.40	1.56
H_2O+	0.47	0.10	0.21	0.22	0.21	0.15	0.38	0.21	0.18	0.12
H_2O-	0.40	0.45	0.48	0.04	0.32	0.60	0.75	0.65	0.34	0.24
CO_2	nil	nil	nil	nil	tr.	nil	nil	tr.	nil	nil
TiO_2	0.65	0.50	0.60	0.40	0.20	0.30	0.50	0.95	1.14	0.75
P_2O_5	0.05	tr.	tr.	0.01	0.02	tr.	tr.	tr.	tr.	0.10
MnO	0.12	0.16	0.21	0.15	0.10	0.30	0.09	0.17	0.33	0.10
Total	99.77	99.43	99.92	99.99	99.71	100.15	99.49	99.54	99.73	100.57

1. Chilled base, at contact with the underlying Ross Sandstone, 1,025 feet below the summit and 3,125 feet above sea-level on the Pinnacles Road, Mount Wellington.

2. Chilled base, 150 feet above the lower contact and 3,275 feet above sea-level on the Pinnacles Road, Mount Wellington.

3. Magnesia-rich stage, from the lower rotunda, 275 feet above the lower contact and 3,400 feet above sea-level on the Pinnacles Road, Mount Wellington.

4. Specimen 25 feet above the lower rotunda, 300 feet above the lower contact and 3,425 feet above sea-level on the Pinnacles Road, Mount Wellington.

5. Feldspar-enriched stage, 325 feet above the lower contact and 3,450 feet above sea-level on the Pinnacles Road, Mount Wellington.

6. Iron-pyroxene enriched stage, 425 feet above the lower contact and 3,550 feet above sea-level on the Pinnacles Road, Mount Wellington.

7. Specimen, 575 feet above the lower contact and 3,700 feet above sea-level on the Pinnacles Road, Mount Wellington.

8. Specimen, 725 feet above the lower contact and 3,850 feet above sea-level on the Pinnacles Road, Mount Wellington.

9. Specimen, 875 feet above the lower contact and 4,000 feet above sea-level on the Pinnacles Road, Mount Wellington.

10. Dioritic stage, summit of Mount Wellington, at 4,150 feet above sea-level and 1,025 feet above the lower contact.

A. B. Edwards, analyst.

original sedimentary cover and a thickness of about 200 feet of the sill itself have been eroded. This can be gauged from the analytical data obtained at Mount Nelson (see below, p. 470), where the top of another sill is preserved in a downfaulted block. Specimens were taken at 50-foot vertical intervals from top to bottom of the sill and

later at 25-foot vertical intervals between 3,400 feet and 3,550 feet above sea-level (between the lower and upper rotundas on the Pinnacles road), where the change in composition of the sill is most marked. Ten chemical analyses were made from these specimens (Table 4). They show that extensive differentiation occurred within the sill after its emplacement.

Oxide profiles.—Interpretation of the analyses is aided by plotting what may be termed "oxide profiles," in which the several oxide

FIG. 7.—The cliffs of Tasman Island, a dissected dolerite sill more than 1,000 feet thick. (Photo by D. J. Mahony.)

constituents of the analyses are plotted against elevation above sea-level (Fig. 8). A marked concentration of MgO occurs immediately above the chilled base of the sill and is sharply demarcated from the upper, more slowly cooled, part. The chilled bottom of the sill has served as a floor on which early-formed magnesia-rich pyroxenes have accumulated by sinking down from the more slowly cooled part of the sill. The shape of the alumina profile shows that this involved a reciprocal displacement upward of the remaining liquid portion of the sill as a whole. The analysis of rock from the summit of Mount Wellington (Table 4, Analysis No. 10) corresponds closely to that of the most acid phase found in the downfaulted block on Mount Nelson (Table 5, Analysis No. 9), which still retains its

upper chilled zone and some of its sedimentary cap. It is assumed, therefore, that not far above the present summit of Mount Wellington, the MgO profile of the sill would follow a course like that outlined by the broken line of Figure 9 and return to an MgO content identical with that of the chilled base. This suggests that the original thickness of the Mount Wellington sill was between 1,200 and 1,300 feet.

The FeO profile shows that, while some iron accompanied the magnesium in the early-formed pyroxenes and accumulated in the lower part of the sill, much of the iron remained in the residual melt

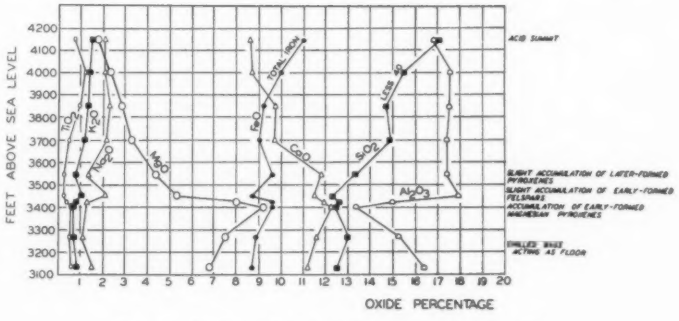


FIG. 8.—Oxide profiles of the Mount Wellington sill

and crystallized in the later-formed pyroxenes, which became progressively richer in iron and poorer in magnesium as differentiation progressed. The slight deviation in the profile about 150 feet above the top of the magnesia layer (Fig. 8) finds a counterpart in the FeO profile of the Mount Nelson sill (Fig. 9) and appears to mark a slight accumulation of iron-rich pyroxene at this level. Most of the later-formed pyroxene is ophitically intergrown with the plagioclase and has been unable to gravitate differentially. The distinct rise in iron content of the more acid top of the sill must, therefore, indicate a general sinking of interlocked plagioclase and pyroxene with a displacement upward of a residual melt still richer in iron.

The TiO_2 profile shows that the titanium behaved in the same way as the iron.

The CaO profile and the Al_2O_3 profile bear out these conclusions. The CaO profile portrays increasing lime as one descends the sill to the magnesia-rich layer, while the Al_2O_3 profile, which indicates the

TABLE 5

ANALYSES OF MOUNT NELSON DOLERITE SILL

	1	2	3	4	5	6	7	8	9	10
SiO_2	52.70	52.20	52.35	52.60	52.35	52.60	53.80	55.70	56.95	52.60
Al_2O_3	14.40	15.25	16.34	15.12	17.68	17.68	17.52	17.53	16.80	16.80
Fe_2O_3	0.35	0.45	0.67	0.53	0.91	0.20	2.15	0.23	0.06	0.08
FeO	8.33	7.78	7.33	8.13	7.37	9.36	7.06	9.21	9.94	8.64
MgO	9.74	8.54	6.30	8.89	5.72	4.28	3.80	2.79	1.63	6.76
CaO	12.40	12.75	11.90	12.95	11.80	10.85	10.55	9.50	8.75	11.60
Na_2O	0.84	1.18	1.88	0.94	1.33	1.67	1.86	2.29	2.17	1.34
K_2O	0.42	0.76	0.82	0.48	0.87	0.93	1.14	1.33	1.88	0.80
H_2O+	0.12	0.21	0.31	0.12	0.45	1.05	0.32	0.12	0.21	0.18
H_2O-	0.67	0.45	1.32	0.38	0.31	0.30	0.54	0.56	0.55	0.31
CO_2	nil	nil	tr.	nil	nil	nil	nil	nil	nil	nil
TiO_2	0.15	0.32	0.35	0.12	0.38	0.52	0.58	0.85	1.00	0.48
P_2O_5	tr.	0.02	0.01	0.05	0.02	tr.	tr.	tr.	0.10	tr.
MnO	0.15	0.16	0.17	0.20	0.12	0.18	0.10	0.12	0.11	0.28
Total	100.27	100.07	99.75	100.51	99.31	99.62	99.42	100.23	100.09	99.87

1. From 150 feet above sea-level at reference 083.683, Lord's Gully, Mount Nelson.
2. From 200 feet above sea-level on the line 076.667–083.683, Lord's Gully, Mount Nelson.
3. From 300 feet above sea-level on the line 076.667–083.683, Lord's Gully, Mount Nelson.
4. From 350 feet above sea-level on the line 076.667–083.683, Lord's Gully, Mount Nelson.
5. From 500 feet above sea-level on the line 076.667–083.683, Lord's Gully, Mount Nelson.
6. From 700 feet above sea-level on the line 076.667–083.683, Lord's Gully, Mount Nelson.
7. From 850 feet above sea-level on the line 076.667–083.683, Lord's Gully, Mount Nelson.
8. From 950 feet above sea-level from line 076.667 on the Mount Nelson signal station road, at the head of Lord's Gully, Mount Nelson.
9. From 450-foot level on Proctor's Road on the downfaulted block, northwest side of Mount Nelson.
10. From the chilled top, at 650 feet above sea-level on Proctor's Road at the contact with the Permo-Carboniferous sediments on the downfaulted block, northwest side of Mount Nelson.

A. B. Edwards, analyst.

feldspar content, reflects the uniform percentage of the Al_2O_3 in the upper and central parts of the sill and the sudden decrease as the magnesia "floor" is approached. This means that some CaO pyroxene accompanies the MgO pyroxene, though the latter preponder-

ates. The CaO profile also shows a similar deviation at the same point as the FeO profile, without any corresponding change in the Al_2O_3 profile, suggesting that some later-formed CaO pyroxene accumulated with the FeO pyroxene. This is confirmed by the petrography of the sill. Between this level and the magnesia-rich layer, the bulges of the Na_2O, K_2O, and Al_2O_3 profiles, combined with the depression in the FeO profile, point to a slight accumulation of plagioclase at about 50 feet above the magnesia floor.

The Na_2O content of the sill reaches a maximum about 300 feet below the present summit and then declines in the higher levels,

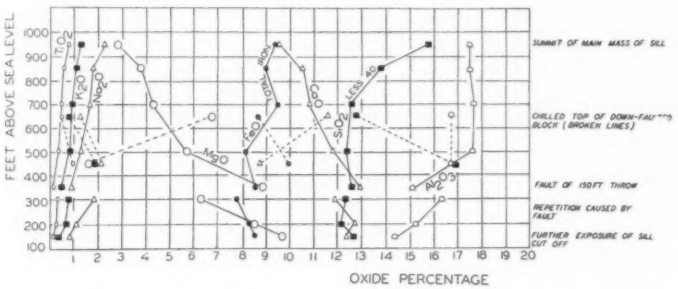


FIG. 9.—Oxide profiles of the Mount Nelson sill, showing the repetition of the curves due to faulting.

whereas the K_2O continues to increase throughout. If this process had been carried further, it must ultimately have given rise to a quartz-monzonite or granitic residuum. A slower rate of cooling, without the formation of ophitic intergrowths between the later-formed pyroxenes and the plagioclase, would have resulted in the development of alternating layers of pyroxene-rich and plagioclase-rich rock in the sill, above the magnesia floor.

2. MOUNT NELSON

Mount Nelson, which lies to the south of the city of Hobart, consists of a similar dolerite sill about 1,000 feet thick, intruded into Permo-Carboniferous sediments. Parts of the sill have been displaced relative to one another by normal faults of Cainozoic age and, on one of the downfaulted blocks on the northwestern flank of

the mountain, the chilled top and portion of the sedimentary roof is preserved. The chilled base is masked by hillwash and possibly by faulting. A suite of specimens collected by Dr. A. N. Lewis, of Hobart, comprises specimens from the downfaulted block with the chilled top and a series taken at 50-foot vertical intervals along a line more or less at right angles to the contour from the 950-foot contour on the road to Mount Nelson signal station to the 150-foot level, below which further outcrop is masked by hillwash. This line runs down Lord's Gully, from reference 076.667, about 2,000 yards from the signal station, to reference 083.683, on Lewis and Murray's inch-to-the-mile map of Hobart (1936) and crosses the Sandy Bay fault.

Ten chemical analyses were made from these specimens (Table 5). They show that the Nelson still underwent differentiation *in situ* in a manner closely comparable with the Mount Wellington sill.

Oxide profiles.—The oxide profiles (Fig. 9) are complicated by the faulting. The profiles for part of the downfaulted block are shown by broken lines, while those for the main block of the mountain are shown by continuous lines. The Sandy Bay fault, which is not detectable in the field where it is wholly within the dolerite, is readily discerned from the profiles by the repetition that results on crossing the fault. This shows, moreover, that the fault caused a downthrow of from 150 to 170 feet on its northeast (seaward) side.

Comparison of Figure 9 with Figure 8 shows that, as at Mount Wellington, the major factor in the differentiation of the sill was the sinking of early-formed magnesia-rich pyroxenes and the concentration of iron in the upwardly displaced residual melt. The gradient of the MgO profile is flatter in the Mount Nelson graph, and the top of the magnesia-rich layer is not so sharply marked, suggesting that the separation of the MgO pyroxene was not so complete as in the case of Mount Wellington. On the other hand, the amount of MgO in the magnesia layer is somewhat greater.

As in the Mount Wellington graph, the FeO profile shows a small deviation, suggesting a slight accumulation of FeO-rich pyroxene above the MgO-rich layer. The deviation occurs considerably higher above this layer than for Mount Wellington, which conforms with the steeper gradient of the MgO curve. The Na_2O again reaches a

maximum and begins to decrease before the most acid layer of the sill is reached, while the K_2O continues to increase throughout. The TiO_2, as at Mount Wellington, is associated with the FeO.

TABLE 6

	Ben Lomond				Mount Sedgwick			Cradle Mountain	
	1	2	3	4	5	6	7	8	9
SiO_2	53.00	52.35	52.45	53.45	52.54	52.25	52.55	52.60	51.95
Al_2O_3	17.12	14.41	14.37	17.79	16.89	13.05	17.60	15.06	13.10
Fe_2O_3	1.02	1.96	1.43	2.20	0.27	0.05	1.30	2.02	1.86
FeO	7.27	7.42	6.98	7.92	8.09	8.27	7.84	7.92	6.61
MgO	6.25	7.75	9.74	3.87	6.36	10.41	4.99	7.33	10.19
CaO	11.30	11.70	12.50	10.50	11.10	12.20	11.85	11.90	12.30
Na_2O	1.30	1.36	1.00	1.79	1.82	0.92	1.32	0.95	1.08
K_2O	0.78	0.98	0.78	0.91	0.83	0.64	0.80	0.72	0.76
H_2O+	0.38	0.56	0.25	0.25	0.21	0.32	0.36	0.25	0.43
H_2O-	0.58	1.03	0.58	0.61	0.76	0.85	0.44	0.63	0.68
CO_2	nil	nil	nil	nil	nil	nil	nil	nil	nil
TiO_2	0.62	0.42	0.15	0.49	0.65	0.50	0.31	0.36	0.40
P_2O_5	tr.	tr.	tr.	tr.	tr.	tr.	0.05	tr.	0.01
MnO	0.16	0.12	0.13	0.07	0.18	0.05	0.19	0.18	0.17
Total	99.78	100.06	100.36	99.85	99.70	99.51	99.60	99.92	99.54

1. Ben Lomond, a few feet above the chilled base above Storey's Creek township.
2. Ben Lomond, about 150 feet above the chilled base above Storey's Creek township.
3. Ben Lomond, about 250 feet above the chilled base above Storey's Creek township.
4. Ben Lomond, summit of the columns, about 450 feet above the chilled base above Storey's Creek township.
5. Mount Sedgwick, a few feet above the chilled base on the eastern side.
6. Mount Sedgwick, about 200 feet above the chilled base on the eastern side.
7. Mount Sedgwick, summit, about 300 feet above the chilled base.
8. Cradle Mountain, about the top of the screes (about 100 ft. above the chilled base).
9. Cradle Mountain, about midway up the dolerite cap (about 300 ft. above the chilled base).

A. B. Edwards, analyst.

3. OTHER SILLS AND LACCOLITHS

Further collections were made at Ben Lomond, Mount Nicholas, and Mount Barrow, in the northeast of Tasmania; at Cradle Mountain and Barn Bluff, in the northwest; at Mount Sedgwick in the west; and at Mount Olympus (Lake St. Clair) and along the scarp of the Western Mountains, between Tunbridge and Interlaken (Lake

Sorell) in the central part of the island. Of these dolerite masses, Ben Lomond (Fig. 2) is the remains of a large sill;[27] Mount Nicholas is the remnant of a smaller sill;[28] Mount Barrow is the remains of a sill or of a laccolith. Cradle Mountain (Fig. 5) and Barn Bluff are two remnants of a dissected sill;[29] Mount Olympus is a remnant of another large sill, and the dolerite forming the Western Mountains is either an enormous sill or "an asymmetric transgressive igneous mass of a general laccolithic type."[30] Mount Sedgwick is either an outlier from the large Eildon sill[31] or the remains of a laccolith.[32]

Examination of thin sections of the rocks from these various localities shows that they do not differ from one another or from those already described in any essential. Sufficient chemical analyses were made from Ben Lomond, Mount Sedgwick, and Cradle Mountain (Table 6) to establish that at each of these localities differentiation of the dolerite had taken place after emplacement with the formation of an MgO-rich layer above the chilled base, with a corresponding enrichment of the residual melt in iron.

II. THE DOLERITE DIKES

4. GUNNING'S SUGARLOAF

Gunning's Sugarloaf, near Campania (Figs. 10, 11) in the southeast part of the Midlands district, was selected as a typical example of the large dikelike intrusions of dolerite characteristic of this part of Tasmania. The dolerite body, which has been mapped by P. B. Nye[33] forms a mountain 6 miles long by 1 mile broad, rising to a height of 1,800 feet above sea-level. The long axis of the mountain trends about 30° E. of north. The dolerite is intrusive into Trias-Jura sandstones, which dip at 5° west. Chilled margins of the dolerite

[27] "The Coal Resources of Tasmania," *op. cit.*, p. 205.

[28] *Ibid.*, pp. 34–35.

[29] Benson, "Notes on the Geology . . . . ," *op. cit.*

[30] Loftus Hills, "The Tasmanian Diabase and Its Relation to the Coal-bearing Formations," in "The Coal Resources of Tasmania," *Mineral Resources, Geol. Surv. Tasmania*, No. 7 (1922), p. 8.

[31] Twelvetrees, "Outlines of . . . . ," *op. cit.*, p. 155.

[32] Edwards, "On a Remnant . . . . ," *op. cit.*

[33] "The Underground Water Resources of the Jericho-Richmond-Bridgewater Area," *Underground Water Supply Papers, Geol. Surv. Tasmania*, No. 2 (1922).

FIG. 10.—End on view of Gunning's Sugarloaf (1,800 ft.), a dikelike intrusive mass of dolerite, near Campania, in the Midlands, looking northwest. (Photo by Q. J. Henderson.)

FIG. 11.—Longitudinal view of Gunning's Sugarloaf, seen from across the Coal River. (Photo by Q. J. Henderson.)

occur at about 500 feet above sea-level on either side of the ridge.

A suite of specimens was collected by Dr. A. N. Lewis, of Hobart, at 50-foot vertical intervals on a line across the summit of the Sugarloaf, running normal to its long axis and extending from chilled margin to chilled margin. Seven chemical analyses were made from the specimens taken between the western chilled margin and the sum-

TABLE 7

	1	2	3	4	5	6	7
SiO_2	52.75	52.50	52.40	52.75	53.55	54.10	54.74
Al_2O_3	15.89	15.79	16.82	17.10	17.22	18.49	19.56
Fe_2O_3	0.81	1.40	0.31	1.61	0.76	0.91	0.27
FeO	8.50	7.49	8.92	7.12	7.78	8.22	8.13
MgO	6.54	6.14	5.67	5.61	4.65	3.29	2.34
CaO	12.30	12.70	11.90	12.30	11.40	11.20	10.98
Na_2O	1.64	1.54	1.41	1.49	1.29	1.69	2.15
K_2O	1.04	0.78	0.83	0.99	1.28	1.25	1.47
H_2O+	0.18	0.19	0.12	0.15	0.42	0.05	0.23
H_2O-	0.20	0.19	0.12	0.15	0.42	0.34	0.31
CO_2	nil	nil	nil	nil	nil	nil	nil
TiO_2	0.50	0.44	0.46	0.60	0.70	0.82	0.66
P_2O_5	tr.	0.01	0.02	0.01	0.01	0.01	0.05
MnO	0.08	0.05	0.11	0.12	0.11	0.18	0.15
Total	100.44	99.41	99.45	100.25	99.47	100.55	100.04

1. Chilled edge at 475 feet above sea-level on the south side of Gunning's Sugarloaf.
2. From 710 feet above sea-level on the south side of Gunning's Sugarloaf.
3. From 850 feet above sea-level on the south side of Gunning's Sugarloaf.
4. From 1,000 feet above sea-level on the south side of Gunning's Sugarloaf.
5. From 1,250 feet above sea-level on the south side of Gunning's Sugarloaf.
6. From 1,550 feet above sea-level on the south side of Gunning's Sugarloaf.
7. Summit of Gunning's Sugarloaf at 1,800 feet above sea-level and about the center of the dike.

A. B. Edwards, analyst.

mit (Table 7). In view of the symmetrical cross section of the mountain, these were assumed to be representative of the two flanks.

Oxide profiles.—The oxide profiles constructed from these analyses (Fig. 12) differ from those obtained for the dolerite sills in certain essential features. Within the width of the chilled margin (wall) little or no differentiation has occurred, but toward the center of the dike a progressively greater sinking of pyroxene has occurred, leaving the central part enriched in feldspathic constituents. Basic pla-

gioclase has also tended to sink in the more slowly cooled central portion, since the Na_2O and K_2O contents of the dike increase toward its center.

III. PEGMATITIC SEGREGATIONS

Veins of unusually coarse-grained dolerite akin to pegmatite occur in association with the normal dolerite. One such vein was observed in the eastern face of the Mount Sedgwick laccolith, in western

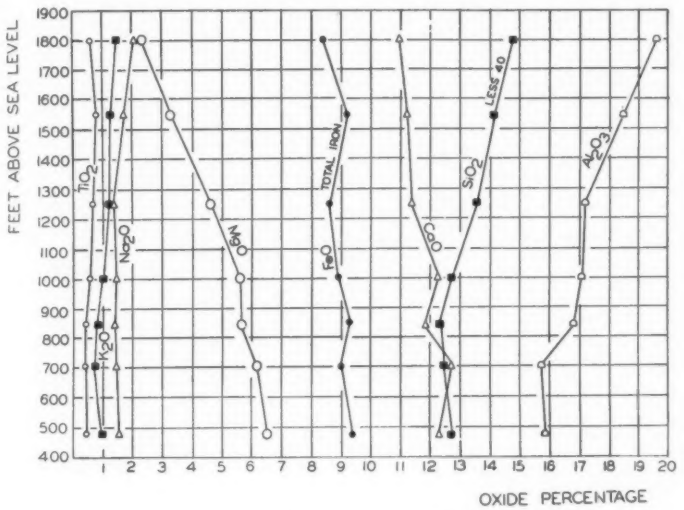


FIG. 12.—Oxide profiles of the Gunning's Sugarloaf dikelike mass

Tasmania, about 100 feet above its base, where it occurred as a lens-like body about 2 inches thick, extending for several feet along a vertical joint plane. The transition from the coarse-grained rock to the finer-grained dolerite was sharply defined, but the transition from one variety to the other was continuous. Similar coarse-grained rock was found on Mount Barrow, in northeastern Tasmania, where it formed a thin facing to a large boulder of fine-grained dolerite. Here, again, the transition from coarse-grained to fine-grained rock was sharp but continuous.

Dr. A. N. Lewis has found further examples of these coarse

pegmatitic rocks in the vicinity of Cambridge and in a road cutting on the Myrtle Gully road, about two miles south of Collins Vale, in eastern Tasmania. At each of these localities the pegmatite occurs as a series of veins an inch or two wide and several feet long, filling what are presumed to be horizontal joint planes in fine-grained dolerite. The pyroxene crystals in the pegmatite tend to orient themselves comb-fashion along the dolerite walls. The boundary line between the coarse- and fine-grained rocks is sharply defined but shows irregular salients, and the transition from one rock to the other, though sudden, is continuous.

TABLE 8

	1	2	3
SiO_2	53.10	55.45	52.50
Al_2O_3	14.13	12.10	16.65
Fe_2O_3	2.14	4.69	0.40
FeO	8.16	8.92	8.82
MgO	6.66	5.68	6.68
CaO	11.50	9.80	11.80
Na_2O	1.28	1.32	1.35
K_2O	1.68	1.26	0.80
H_2O+	0.31	0.05	0.31
H_2O-	0.72	0.42	0.42
CO_2	tr.	nil	nil
TiO_2	0.50	0.80	0.60
P_2O_5	0.05	0.01	tr.
MnO	0.17	0.18	0.15
Total	100.40	100.68	100.48

1. Coarse vein, in a vertical joint plane about 100 feet above the chilled base, Mount Sedgwick.

2. Coarse vein, in a horizontal joint plane, cutting on the Myrtle Gully road, two miles south of Collins Vale.

3. Normal fine-grained dolerite, three inches away from the coarse vein, cutting in the Myrtle Gully road, two miles south of Collins Vale.

A. B. Edwards, analyst.

Analyses of the pegmatites from Mount Sedgwick and from the Myrtle Gully road (Table 8) show that they are of variable composition and that they differ in composition both from the host-rock and from the normal differentiation products of the dolerite. They are rather more acid and contain distinctly more K_2O and less MgO and Al_2O_3 than their host-rocks. They also contain a notable amount of

ferric iron. These features of their chemical composition are reflected in their mineralogical composition.

The pegmatites consist of flattened and elongated prisms of pyroxene and coarse laths of basic plagioclase, set in an abundant mesostasis of cloudy feldspathic material, which may constitute as much as one-third of the rock. Two series of monoclinic pyroxenes are present: augite of large optic axial angle and pigeonite of low optic axial angle; and both sometimes show partial alteration to chlorite. Individual pyroxene grains are as large as $30 \times 3 \times 1$ mm., but the average grain size is much smaller, and most crystals do not exceed 5 mm. in length. The mesostasis consists of long slender rods of plagioclase in a base of another feldspar, which appears to be orthoclase. The plagioclase rods sometimes form radiating or plumose sheafs: sometimes they occur in parallel groups of three or four rods, several groups crossing one another at various angles; and sometimes they form an irregular network. Associated with this feldspathic intergrowth are numerous patches of micrographically intergrown quartz and orthoclase(?) and coarser, irregular, grains of quartz. In the Mount Barrow pegmatite and the Mount Sedgwick pegmatite, minute needles of pyroxene occur in the mesostasis, and, in places, clusters of small elongated grains of pyroxene with parallel extinction occur. These growths represent the early stages in the development of ophitic intergrowths of pyroxene and plagioclase, and various transitional stages between this stage and the completed intergrowth can be seen. The iron ore is present as numerous crystals of magnetite. They are usually rectangular in outline and may be as large as 1×0.5 mm. In some sections the iron ores also form strings of granules in the mesostasis and are concentrated as fringes to pyroxene crystals, having been precipitated during the marginal alteration of the pyroxene to chlorite. Occasionally, small flakes of brown biotite are present with the chlorite, and in the Mount Sedgwick rock the mesostasis contains a number of small, late-formed prismatic crystals of emerald-green hornblende.

Browne[34] has drawn attention to the close similarity between a coarse-grained segregation vein of this character, exposed in a cutting in the Domain, in Hobart, and a specimen of coarse-grained

[34] *Op. cit.*, p. 251.

dolerite from the Horn Bluff, King George Land, Antarctica. These pegmatites also resemble the dolerite pegmatites of other dolerite bodies like the Whin sill[35] and the Palisade sill.[36] They seem best explained as small quantities of residual magma which were trapped in the more rapidly solidified portions of the dolerite masses, so that their volatiles could not escape into the still liquid cores of the intrusions, and which were later squeezed into joint planes that formed as the still hot rock began to contract. As indicated by Tomkeieff,[37] retention of the volatiles would lead to coarser crystallization at a slightly lower temperature than that at which crystallization occurred in the main body of the rock and might reduce the affinity between the iron oxides and silica, leading to an increased formation of iron ore and a liberation of free silica, which would enter into micrographic growths with alkali feldspar or might even appear, as in the Tasmanian pegmatites, as small allotriomorphic crystals of quartz. In the final stages of solidification the concentrated volatiles, especially the water, might be expected to convert the pyroxene into chlorite to greater or less degree, with simultaneous precipitation of iron ores, and might occasionally lead to the formation of fairly sodic hornblende in the mesostasis. Similar trends might be expected in the more acid phases of the sills, where the volatiles tended to be concentrated, and this is found to be so.

[35] S. I. Tomkeieff, "A Contribution to the Petrology of the Whin Sill," *Min. Mag.*, Vol. XXII (1929), pp. 102–9.

[36] Walker, *op. cit.*, p. 1095.

[37] *Op. cit.*, p. 118.

[*To be continued*]

MULTIPLE PLEISTOCENE GLACIATION ON SAN FRANCISCO MOUNTAIN, ARIZONA

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ABSTRACT

Evidence of three distinct glaciations, tentatively dated as later Wisconsin, Iowan-Wisconsin, and Illinoian, is recognized on San Francisco Mountain. Wisconsin snow lines and the southernmost extent of mountain glaciation in the United States are treated briefly, and features produced by frost action are described. The relation between glaciation and volcanic eruptions is discussed.

INTRODUCTION

The glaciation of San Francisco Mountain was first reported by W. W. Atwood[1] and has subsequently been described by others.[2] All the evidence previously reported is related to the youngest of the three glaciations described in this article, the two earlier glaciations having been unrecognized heretofore. Field investigations were made during a month's stay in the area during June and July of 1940.

PHYSICAL SETTING

San Francisco Mountain lies near the southern margin of the Coconino Plateau in north-central Arizona at 35°20′ N. and 111°40′ W. The city of Flagstaff on the Santa Fe Railroad is 9 miles due south, and the Grand Canyon is 60 miles northwest. The mountain and the surrounding country are shown on the Flagstaff quadrangle.

San Francisco Mountain is a composite volcanic cone rising 5,500 feet above the level of the plateau. It is breached on the northeast

[1] Glaciation of San Francisco Mountain, Arizona," *Jour. Geol.*, Vol. XIII (1905), pp. 276–79.

[2] D. W. Johnson, "Report on the Geological Excursion through New Mexico, Arizona, and Utah, Summer of 1906," *Technology Quart.*, Vol. XIX (1906), pp. 408–15; "The Southernmost Glaciation in the United States," *Science*, Vol. XXXI (1910), pp. 218–20; H. H. Robinson, "The San Franciscan Volcanic Field, Arizona," *U.S. Geol. Surv. Prof. Paper 76* (1913), pp. 31–33; H. S. Colton, "San Francisco Peaks," *Museum Northern Arizona, Museum Notes*, Vol. III, No. 3 (1930), p. 2; H. S. Colton and F. C. Baxter, "Days in the Painted Desert and the San Francisco Mountains," *Northern Arizona Soc. Sci. Art. Bull. 2* (1932), p. 11.

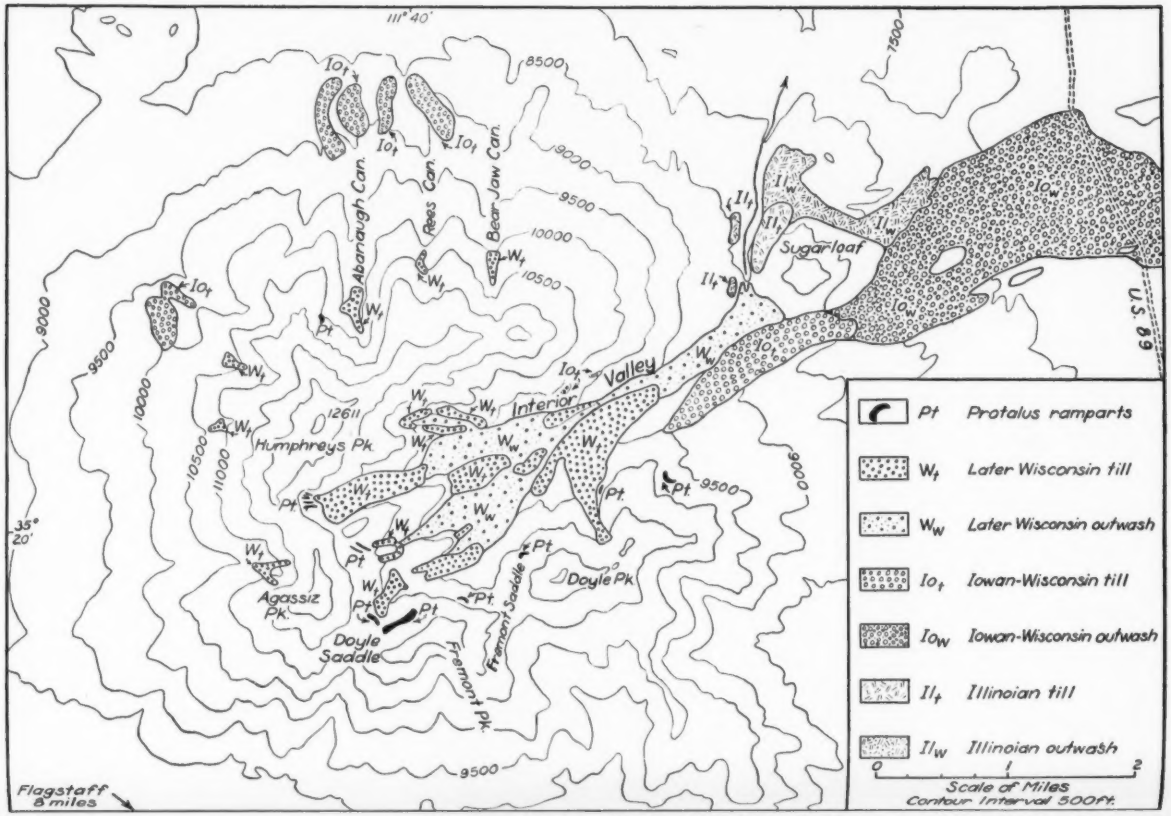

FIG. 1.—Glacial deposits on San Francisco Mountain, Arizona

by a large canyon locally known as the "Interior Valley," which contained the largest glaciers of the area. Humphreys Peak (San Francisco Peak), at 12,611 feet, is the highest of the several lofty peaks near the head of the Interior Valley. The timber line is close to 11,500 feet.

The cone consists of lava and pyroclastics erupted early in the

FIG. 2.—View eastward from head of north branch of Interior Valley showing back of later Wisconsin recessional moraine in foreground, later Wisconsin outwash in valley below, Iowan-Wisconsin lateral moraine (*AA*), and late Pleistocene basaltic cones in mid-distance.

Pleistocene.[3] Common lithologic types are latite, pyroxene dacite, hornblende dacite, rhyolite, and andesite. A number of younger basaltic flows and cones[4] cluster around the mountain (Fig. 2).

DESCRIPTION OF GLACIATION

Stages of mountain glaciation are usually given local names, but in this instance the glaciers were so few and so small that applica-

[3] Robinson, *op. cit.*, pp. 42–52, 95.

[4] Colton, "The Basaltic Cinder Cones and Lava Flows of the San Francisco Mountain Volcanic Field, Arizona," *Museum Northern Arizona Bull. 10* (1937), pp. 1–50.

tion of special names does not seem justified. For this reason, names of the apparently correlative stages or substages in the Midwest are adopted without meaning to imply that the correlations are established beyond doubt.

ILLINOIAN GLACIATION

The evidence for this glaciation consists of remnants of lateral moraines and outwash deposits at the mouth of the Interior Valley near Sugarloaf (Fig. 1). Here, as elsewhere in the western mountains, earlier glaciers appear to have been more extensive than the later ones.[5] An Illinoian glacier at least $4\frac{3}{4}$ miles long with a terminus at 8,500 feet occupied the Interior Valley. The ice was 1,000 feet thick, and possibly it spilled through Fremont Saddle and down the south slope of the mountain, though the evidence is not convincing.

Illinoian moraines.—At Sugarloaf remnants of till are perched on bedrock shoulders 350–370 feet above the floor of the Interior Valley (Fig. 1). The till remnant on the east has the form of a lateral moraine (Fig. 3) and is much the larger and better preserved, because its position renders it less vulnerable to erosion than the till on the west, which has been greatly eroded by runoff from an extensive mountain slope above.

Distinguishing features of these till deposits are: the large amount of rhyolite contained, the scarcity of boulders on the surface, and their topographic position, all of which are in marked contrast to the next younger (Iowan-Wisconsin) moraines a short distance up the Interior Valley. Along the crest of the eastern moraine the frequency of boulders larger than 1 foot in diameter is about one per 100 feet. These boulders consist of relatively dense, resistant rock types, chiefly andesite and dacite. The rhyolitic material in the Illinoian drift could have come only from Doyle Saddle or Sugar-

[5] Atwood, "Glaciation of the Uinta and Wasatch Mountains," *U.S. Geol. Surv. Prof. Paper 61* (1909), p. 67; Eliot Blackwelder, "Post-Cretaceous History of the Mountains of Central Western Wyoming," *Jour. Geol.*, Vol. XXIII (1915), p. 333; "Pleistocene Glaciation in the Sierra Nevada and Basin Ranges," *Bull. Geol. Soc. Amer.*, Vol. XLII (1931), p. 872; W. W. Atwood and Kirtley F. Mather, "Physiography and Quaternary Geology of the San Juan Mountains, Colorado," *U.S. Geol. Surv. Prof. Paper 166* (1932), pp. 28, 82; F. M. Fryxell, "Glacial Features of Jackson Hole, Wyoming," *Augustana Library Pub. 13* (1930), p. 35.

loaf.[6] Undoubtedly, much of the rhyolite did come from Sugarloaf, but patches of rhyolitic detritus on the north slope of Doyle Peak above the limit of the Wisconsin glaciers indicate that some of it also came from Doyle Saddle. It seems that the amount of rhyolite in Doyle Saddle was considerably larger in Illinoian time than at present.

Not the least significant feature of the Illinoian moraines at Sugarloaf is their topographic position. They rest on a bedrock surface which is an ancient valley floor. The present stream has cut a

FIG. 3.—View east-northeast to west slope of Sugarloaf showing Illinoian lateral moraine (*AA*) at its west base.

narrow, steep-walled gorge 370 feet deep into this rock floor, leaving the moraines on bedrock shoulders high above. The fact that these moraines were deposited on the floor of the ancient valley and not on its walls suggests that the terminus of the glacier was not much farther down the valley.

Illinoian outwash deposits.—The east lateral moraine at Sugarloaf passes northward into a gently sloping, smooth-surfaced gravel deposit resting on the same bedrock floor. This gravel is also dominantly rhyolitic and is believed to constitute a valley train deposited by Illinoian meltwater flowing in the direction of the present Interior Valley drainage. Part of the Illinoian meltwater also escaped through a low saddle between Sugarloaf and the hill next north, to

[6] Robinson, *op. cit.*, p. 47, Pl. V.

form a large deposit northeast of Sugarloaf and outside of the Interior Valley drainage (Fig. 1). These deposits are well exposed in gravel pits, where they are at least 100 feet thick and consist predominantly of rhyolite and minor amounts of andesite, dacite, and pyroclastics of the types outcropping near the head of the Interior Valley. Sand-sized particles predominate, but layers of pebbly gravel and a few boulders up to 1½ feet in diameter are also present. Foreset bedding, scour channels, and crude sorting and bedding throughout suggest rapid construction of a fanlike deposit under floodwater conditions. The lithologic constitution of this material, the structural evidences of its mode of origin, and the fact that it can be traced into a remnant of the Illinoian valley train indicate deposition by Illinoian meltwater. Much of the rhyolite was probably picked up by the meltwater from the slopes and area surrounding Sugarloaf and is not truly glaciofluvial in the strictest sense.

Age and correlation.—The criteria used in determining the age of mountain glacial deposits have been discussed elsewhere.[7] Of these criteria the topographic form and position of the moraines, the composition and relative abundance of boulders on their surfaces, and the amount of dissection of moraines and bedrock have been the most useful here.

The age of the Illinoian deposits is attested by their subdued topographic form, the scarcity of boulders on the surface, the fact that all these boulders are of resistant rock types, and the 370 feet of subsequent bedrock dissection. This last feature is considered the most significant, for a distinguishing feature of nearly all pre-Wisconsin glacial deposits in western mountains is their position at some considerable height above the present streams.[8] Depth of dissection

[7] Blackwelder, "Pleistocene Glaciation in the Sierra Nevada . . . . ," *op. cit.*, pp. 870–80; R. P. Sharp, "Pleistocene Glaciation in the Ruby-East Humboldt Range, Northeastern Nevada," *Jour. Geomorph.*, Vol. I (1938), pp. 300–304.

[8] Blackwelder, "Post-Cretaceous History . . . . ," *op. cit.*, p. 322; "Pleistocene Glaciation . . . . ," *op. cit.*, pp. 896, 906; Fryxell, *op. cit.*, pp. 16, 36; Atwood and Mather, *op. cit.*, pp. 44, 50, 53, 60, 108; W. C. Alden, "Physiography and Glacial Geology of Eastern Montana and Adjacent Areas," *U.S. Geol. Surv. Prof. Paper 174* (1932), p. 31; Neil A. Miner, "Evidence of Multiple Glaciation in the Northern Part of Yellowstone National Park," *Jour. Geol.*, Vol. XLV (1937), p. 645; Leland Horberg, "Structural Geology and Physiography of the Teton Pass Area, Wyoming," *Augustana Library Pub. 16* (1938), p. 74.

is dependent upon a variety of factors, of which size and gradient of the stream, nature of bedrock, and diastrophic movements are of prime importance. In this case a small intermittent stream has cut a gorge 370 feet deep into a dense, uniform, relatively resistant andesite. Nowhere in the western mountains has there been reported such a depth of dissection in bedrock since the earliest Wisconsin (Iowan) glaciation except along large powerful rivers where conditions for dissection are much more favorable. This relation, and the fact that the next youngest moraines on San Francisco Mountain are dated as Iowan-Wisconsin with considerable confidence, indicates that the moraines at Sugarloaf are pre-Wisconsin. The glaciation recognized as preceding the early Wisconsin in most western mountains is dated as Illinoian or Kansan, with Kansan favored because of Blackwelder's[9] interpretations. Indisputable proof of Illinoian glaciation in the western mountains has not been presented,[10] although evidence for such a glacial stage is not entirely lacking.[11] The normal procedure, therefore, would be to refer the moraines at Sugarloaf to the Kansan; but, in so far as observed by the writer, the Kansan materials in the Sierras and the San Juans appear much older than the drift at Sugarloaf, being deeply weathered and having lost all traces of glacial topography. The moraine just west of Sugarloaf has distinctly the form of a lateral moraine (Fig. 3), and the depth and degree of weathering do not appear comparable to those of the Kansan deposits. Such marked differences cannot be attributed to climatic influences, for climate and vegetation are reasonably similar in all three areas.

For these reasons the Sugarloaf moraines are tentatively assigned to the Illinoian rather than to the Kansan.

[9] "Pleistocene Glaciation . . . . ," *op. cit.*, p. 918.

[10] Leland Horberg, of the University of Illinois, suggests that glaciers were small or absent in the western mountains during the Illinoian stage because the Cordilleran and Keewatin centers were not particularly active and because the greatest area of accumulation lay farther east in the Labradorian center, as shown by the Illinoian deposits in the Midwest. It is reasonable to expect that the glaciers of the western mountains would show advances and retreats more in sympathy with the Cordilleran and Keewatin centers than with the Labradorian center. This same view with respect to the Illinoian glaciation has been expressed by Antevs in subsequent correspondence.

[11] Blackwelder, "Pleistocene Glaciation . . . . ," *op. cit.*, pp. 895, 918–19.

IOWAN-WISCONSIN GLACIATION

The evidence for this glaciation consists of a number of lateral moraines and a considerable area of outwash. The Interior Valley glacier of this substage extended down to the west slope of Sugarloaf at 8,600 feet. It was fully 4 miles long and at least 600 feet thick. Glaciers approximately 2 miles long also occupied canyons such as Abanaugh and Rees on the northern slope of the mountain (Fig. 1).

Iowan-Wisconsin moraines.—The prominent ridge forming the south side of the Interior Valley below the 9,500-foot contour is a large lateral moraine of the Iowan-Wisconsin glaciation (Fig. 2). The abundance and size of the boulders on the surface of this moraine attract immediate attention. They are as large as 10 feet in diameter and consist of gray, black, white, and reddish dacite and dacite tuff, vesicular and dense andesite, andesitic agglomerate, and pink coarse-grained dacite and latite. Many of these are rock types exposed near the head of the Interior Valley. Small remnants of the complementary lateral moraine are preserved on the northwest wall of the Interior Valley, but this moraine has been largely washed away by water supplied from an extensive mountain slope above. The terminal loop which joined the two laterals across the Interior Valley has been entirely removed by stream erosion. Bulky Iowan-Wisconsin lateral moraines are also preserved along some canyons on the north and northwest flanks of San Francisco Mountain (Fig. 1).

Iowan-Wisconsin outwash.—The largest area of Iowan-Wisconsin outwash recognizable as such is east-northeast of Sugarloaf (Fig. 1). The derivation of this material from a glacial source is shown by its content of striated boulders in a gravel pit 250 yards east of U.S. Highway 89. The general form of the outwash deposit is that of a broad alluvial fan, thicker near its head and thinner outward, where it has been dissected by streams and can be seen to rest on a bedrock surface of moderate relief. The greatest thickness measured was 88 feet at a point approximately 1 mile from the foot of the moraine. Like the moraine, this outwash consists of subangular to rounded cobbles and boulders of lava, tuff, and agglomerate of dacitic, latitic, and andesitic composition. This material could not have been transported through the regular outlet channel of the Interior Valley but must have come from the ice where it surmounted the low southeast

wall of the Interior Valley just south of Sugarloaf. Part of the Illinoian ice may also have passed through this low saddle, but there is no evidence on this point, and the passage of a considerable part of the Iowan-Wisconsin glacier along this route suggests lowering of the ridge and saddle by erosion during the preceding interglacial stage. Outwash materials were probably deposited at this time both at the outlet of the Interior Valley and at the points of debouchment of the glacial streams on the north and northwest flanks of the mountain, although such deposits have not been positively identified at any of these places.

Age and correlation.—The deposits of this glaciation are confidently correlated with those of the Tahoe stage in the Sierra Nevada[12] because of closely corresponding topographic form and position, amount of succeeding erosion, and relation to moraines of the next younger glaciation. Such correlation also implies that this glaciation is the same as the Durango in the San Juans, the Bull Lake in the Middle Rockies, the Lamoille in Nevada, the Blacks Fork in the Uintas, the Wisconsin I or Twin Lakes of Colorado, and other corresponding glacial advances.[13] These correlations are considered reasonably satisfactory because an almost universal relationship in the western mountains is that of large lateral moraines of an earlier glaciation (Iowan-Wisconsin) and well-preserved terminals of a less extensive and considerably younger advance.

The Tahoe and its correlatives are now generally considered Iowan on the strength of Alden's[14] work; and Ernst Antevs[15] also suggests that the Tahoe is Iowan-Wisconsin, because the Cordilleran and Keewatin centers of ice accumulation were active at this time, as

[12] *Ibid.*, pp. 884–95.

[13] Blackwelder, "Post-Cretaceous History . . . . ," *op. cit.*, pp. 325–28; "Pleistocene Glaciation . . . . ," *op. cit.*, p. 918; Atwood and Mather, *op. cit.*, pp. 78–81, 120–33; W. H. Bradley, "Geomorphology of the North Flank of the Uinta Mountains," *U.S. Geol. Surv. Prof. Paper 185-I* (1936), pp. 194–95; Kirk Bryan and L. L. Ray, "Geologic Antiquity of the Lindenmeier Site in Colorado," *Smithsonian Misc. Coll.*, Vol. XCIX (1940), pp. 29, 32; Ray, "Glacial Chronology of the Southern Rocky Mountains," *Bull. Geol. Soc. Amer.*, Vol. LI (1940), p. 1857.

[14] *Op. cit.*

[15] "Climatic Variations during the Last Glaciation in North America," *Bull. Amer. Met. Soc.*, Vol. XIX (1938), p. 172.

shown by the nature of the Iowan drift in the Midwest. Most western glaciologists accept the Iowan as the earliest Wisconsin.

For these reasons the intermediate glaciation on San Francisco Mountain is tentatively dated as Iowan-Wisconsin.

LATER WISCONSIN GLACIATION

This substage is termed "later Wisconsin," rather than "late Wisconsin," because it is not known exactly which substage of the Wisconsin is represented. The evidence consists of terminal, medial, recessional, and ground moraines, outwash deposits, striated bedrock surfaces, and cirques. During this substage the Interior Valley contained a glacier $2\frac{3}{4}$ miles long with a terminus at approximately 9,000 feet. Smaller glaciers occupied the outer canyons on the north and northwest flanks of the mountain.

Later Wisconsin moraines.—The later Wisconsin terminal moraine in the Interior Valley is a large irregular mass of rock debris which completely chokes the valley at 9,000–9,200 feet. This moraine is compound, for it consists of debris left by the Interior Valley glacier and a tributary glacier from the south which joined and partly overrode the trunk glacier. The Interior Valley stream has cut only a narrow slot 50–75 feet deep through the moraine, and the youthfulness of the deposit is further attested by its fresh, essentially unmodified glacial topography, characterized by numerous closed depressions.

The till composing this and other later Wisconsin moraines consists of large angular and subangular blocks of a wide variety of rocks, chiefly andesite, dacite, dacite tuff, and latite, accompanied by small amounts of sandy and gravelly matrix. Weathering is slight, and surface boulders are more abundant than on the Iowan-Wisconsin moraines.

As the ice retreated from the terminal position, it split at once into two parts—a trunk glacier which receded up the Interior Valley, and a tributary glacier which receded into the canyon northeast of Doyle Peak. The trunk glacier soon bifurcated into two approximately equal masses, each confined to one of the two major branches of the Interior Valley. A medial moraine formed between these two branches of the trunk glacier remains as one of the most prominent

and best-preserved moraines on the mountain (Fig. 4). Recession was accompanied by deposition of ground moraine, particularly just behind the terminal position. Further retreat produced a bifurcation of both branches of the glacier into ice streams heading in the four major cirques on the flanks of the high peaks.

FIG. 4.—Later Wisconsin medial moraine in Interior Valley

FIG. 5.—Later Wisconsin recessional moraine in north branch of Interior Valley

About ½ mile below the cirque between Humphreys and Agassiz peaks a moraine completely fills the canyon from wall to wall and rises fully 500 feet above the canyon floor on the downstream side (Fig. 5). This moraine consists of large angular blocks jumbled together in great confusion. Its outer face is covered with trees and bushes and shows minor modification by weathering and erosion. The canyon above the moraine is filled with huge angular boulders. The surface of this boulder accumulation is flush with the top of the

moraine and is extremely rough, consisting of irregular ridges and knobs separated by closed depressions 20–30 feet deep. A few trees grow on some of the ridges, but the lack of soil and vegetation, particularly in the closed depressions, is notable. This boulder filling appears to be younger than the moraine and possibly indicates subsequent readvance or fluctuation of the ice front. The southern branch of the Interior Valley, heading in Doyle Saddle, contains two separate recessional moraines in its upper part (Fig. 1), which might possibly correspond to the moraine and boulder filling just described. A considerable difference in appearance of these deposits, however, prohibits any definite correlation.

During the later Wisconsin glaciation at least six canyons on the northern and western outer slopes of the mountain contained small ice masses which advanced a short distance and built irregular piles of angular blocks, now choking the canyons at elevations between 10,000 and 11,000 feet. None of these deposits constitutes a well-formed moraine, but they are too far from the heads of the canyons to have been formed in any other way. The best-developed member of this group occupies Abanaugh Canyon and was deposited by a glacier about $\frac{1}{2}$ mile long. No traces of the recessional phases recorded in the Interior Valley were observed in these smaller canyons.

Later Wisconsin outwash.—Outwash deposits of the later Wisconsin glaciation have been identified with certainty only in the Interior Valley. Probably some of the water-borne detritus in the other glaciated canyons is also outwash of this age, and much of the material in the fans flanking the mountain at the mouths of glaciated canyons may be of this origin, as suggested by Robinson.[16]

The floor of the Interior Valley between the later Wisconsin terminal moraine and Sugarloaf is covered with a deposit of bouldery gravel which clearly constitutes a valley train. This deposit has been dissected to a depth of 25–30 feet by the modern stream and has not been identified beyond the rock gorge at the mouth of the valley. The ground moraine back of the terminal moraine is largely veneered with outwash, which is best shown in the northern branch of the valley fronting the high recessional moraine previously described. It covers the valley floor from wall to wall and consists of large angu-

[16] *Op. cit.*, p. 32.

lar and subangular blocks of a variety of rocks jumbled together in torrential fashion. The surface of the deposit is scoured by bifurcating and anastomosing channels typical of valley trains.

Miscellaneous features of the later Wisconsin glaciation.—A bedrock surface showing distinct glacial polish and striae was observed at the tip of the bedrock spur separating the northern fork of the Interior Valley from its northern prong. The bedrock is a dense uniform dacite or andesite, and the preservation of the striae and polish is probably due to a fortuitous relation between resistant rock and a protecting till cover only recently removed. Striated bedrock surfaces have been reported elsewhere in the Interior Valley[17] but were not found during this study.

Small deposits of fine sand and silt on the later Wisconsin terminal moraine in the Interior Valley have the aspects of localized accumulations of sandy loess. This is the only feature suggestive of wind action associated with these glaciers.

Age and correlation.—In the Sierra Nevada of California the second major Wisconsin advance has been named the "Tioga,"[18] and the features just described are unhesitatingly correlated with the Tioga on the strength of a similar degree of freshness, position, topography, relation to the earliest Wisconsin features, and depth of dissection. The position of the Tioga and its correlatives within the subdivisions of the type section of the Midwest remains one of the foremost problems of Pleistocene chronology in spite of vigorous attack by many competent glaciologists.

Antevs[19] has advanced the following argument favoring correlation of the Tioga with the late Mankato. The earliest Wisconsin (Iowan) lobe of midwestern United States was formed by accumulation and advance of ice from the combined Cordilleran and Keewatin centers of western Canada. Following this climax, a period of deglaciation occurred in these western centers, but concurrently the Labradorian center became active and supplied ice for the Taze-

[17] Atwood, "Glaciation of San Francisco Mountain . . . . ," *op. cit.*, p. 278; Robinson, *op. cit.*, p. 31.

[18] Blackwelder, "Pleistocene Glaciation . . . . ," *op. cit.*, p. 881.

[19] *Op. cit.*, pp. 172, 176; "The Cochise Culture," *Medallion Papers No. XXIX* (privately printed, 1941), pp. 40–41.

well advance. Successive Wisconsin advances in the Midwest were then nourished from centers of accumulation which moved progressively westward until the Keewatin and Cordilleran areas again became active and supplied ice for the late Mankato advance. It is presumed that the conditions causing accumulation of ice and snow in the Keewatin and Cordilleran centers would favor development and growth of glaciers in the mountains of western United States. This leads to the conclusion that the two major Wisconsin advances in the western mountains should be correlated with the Iowan-Wisconsin and late Mankato of the Midwest.

Bryan and Ray[20] propose a subdivision of the Wisconsin in the southern Rocky Mountains which corresponds in major part to the substages of the Wisconsin in central North America and of the Würm in Europe. It is possible that the Rocky Mountains are far enough east to have been affected by conditions causing advances from the Labradorian center; but it is questionable whether the mountain glaciers would have been progressively shorter, as required by the Bryan-Ray scheme, as the centers of accumulation shifted westward from Labrador.

The Home substage (second major Wisconsin advance) of the southern Rocky Mountains has been dated as Tazewell-Cary.[21] No definite correlation between the Tioga and Home substages has been established, although such a correlation appears reasonable and is apparently favored by Antevs.[22] Under any circumstances, dating of the second major Wisconsin advance in the southern Rockies as Tazewell-Cary and in other western mountains as late Mankato raises a conflict worthy of concerted attack by glaciologists.

Dating of the two major Wisconsin glaciations in the mountains of California, Nevada, and Arizona as Iowan-Wisconsin and late Mankato finds support in the evidence of a considerable time-interval separating these advances. If the time elapsed between the late Mankato and the present is taken as unity, the interval between the Iowan-Wisconsin and the late Mankato was about two units and that between the Tazewell-Cary and the late Mankato about three-

[20] *Op. cit.*, p. 68; Ray, *op. cit.*, p. 1914.

[21] Bryan and Ray, *op. cit.*, pp. 67, 68; Ray, *op. cit.*, pp. 1858–1914.

[22] "The Cochise Culture," *op. cit.*, p. 41.

quarters unity.[23] Of these two figures, the first appears to be more nearly correct, in so far as the length of the interval can be judged from the appearance and condition of moraines. Blackwelder's[24] studies also favor the longer interval.

In view of these considerations, the second Wisconsin advance on San Francisco Mountain is dated simply as later Wisconsin, with a rather late substage of the Wisconsin favored because of the comparatively long interval between the two Wisconsin advances and the explanation for this interval suggested by Antevs.

One, and possibly two, periods of stability or readvance during the recession of the later Wisconsin ice are indicated by the moraines in the Interior Valley. In the Bryan-Ray chronology these would probably be dated as Wisconsin III and Wisconsin IV. Under the Antevs correlation these moraines would be recessional phases of the late Mankato; and in this particular area, at least, the second view appears the more applicable because of the short time-intervals separating all the later Wisconsin moraines.

F. E. Matthes[25] finds evidence of a rebirth and readvance of glaciers about 4,000 years ago in the Sierra Nevada and attributes the extremely fresh moraines at the mouths of many cirques to this "little glaciation." The accumulation of bouldery material behind the first recessional moraine in the canyon heading between Humphreys and Agassiz peaks conforms in fair degree with the fresh moraines in the Sierra Nevada, but there is no way of establishing a definite correlation. It is possible that mountains east of the Sierras did not contain glaciers during the little glaciation because the snow line in such mountains is and has been considerably higher than the snow line in the Sierras.[26]

[23] Antevs, "Climatic Variations . . . . ," *op. cit.*, pp. 173–74.

[24] *Op. cit.*, pp. 880–81.

[25] "The Post-Pleistocene Moraines of the Sierra Nevada," *Annals Assoc. Amer. Geog.*, Vol. VI (1917), pp. 128–29; "Report of Committee on Glaciers, April, 1939," *Trans. Amer. Geophys. Union* (1939), pp. 519–20; "Committee on Glaciers, 1939–40," *Trans. Amer. Geophys. Union* (1940), pp. 398–402; "Rebirth of the Glaciers of the Sierra Nevada during Late Post-Pleistocene Time," *Bull. Geol. Soc. Amer.*, Vol. LII (1941), p. 2030.

[26] Fritz Klute, "Die Bedeutung der Depression der Schneegrenze für eiszeitliche Probleme," *Zeitschrift f. Gletscherk.*, Vol. XVI (1928), p. 75; Herbert Louis, "Die Ver-

In summary, the two Wisconsin advances on San Francisco Mountain are tentatively dated as Iowan-Wisconsin (earliest Wisconsin) and later Wisconsin, with a rather late phase of the Wisconsin being favored. This later Wisconsin substage shows evidence of one, and perhaps two, recessional phases, although there is a possibility that the second phase is actually related to the post-Wisconsin little glaciation of the Sierra Nevada.

EFFECTS OF FROST ACTION

Included for discussion are protalus ramparts, rock streams, stone stripes, terracettes, stone rings, frost-heaved pavement, and modification of land forms.

Protalus ramparts[27] are located on the floors of cirques and at the heads of some unglaciated canyons. The larger ones are shown on the map (Fig. 1). Their form is that of a straight or curved linear embankment, 10–50 feet high, 20–50 feet wide, and 50–400 feet long. They consist of jumbled accumulations of angular blocks, 1–12 feet in diameter, and usually lie within 50–300 feet of the foot of a steep wall or slope from which the boulders have been derived. The form, location, and constitution are explained by accumulation of boulders which have rolled or slid across a snowbank at the foot of the slope. After the snow melts, an embankment is left, separated from the rock wall by a fosse formerly occupied by snow. Intense frost action on the exposed rocky slope and a perennial snowbank are two factors favoring the development of protalus ramparts.

The best-developed group of ramparts lies in the cirque heading between Agassiz and Humphreys peaks. This group is particularly notable because not all its members are of the same age, as can be seen in the photograph, Figure 6, and the sketch, Figure 7. The older ramparts are larger and farther from the wall than the younger set. Their greater age is indicated by subdued outlines, surfaces partly covered by soil, and a considerable vegetative cover, in con-

breitung von Glacialformen im der Westen der Vereinigten Staaten," *Zeitschrift f. Geomorph.*, Vol. II (1928), pp. 221–35.

[27] W. S. Cooper, "Plant Succession in the Mount Robson Region, British Columbia," *Plant World*, Vol. XIX (1916), p. 220; C. H. Behre, "Talus Behavior above Timber in the Rocky Mountains," *Jour. Geol.*, Vol. XLI (1933), p. 630; Bryan, "Geomorphic Processes at High Altitudes," *Geog. Rev.*, Vol. XXIV (1934), pp. 655–56.

trast to the younger ramparts, which are entirely free of soil and vegetation (Fig. 6). The outer ramparts appear to be several times as old as the younger set, which may even be receiving some additions during favorable seasons at present. Obviously, the snowbank was much larger when the older set was formed, and this suggests a somewhat colder period separated from the time during which the younger ramparts formed by a warmer interval when frost action

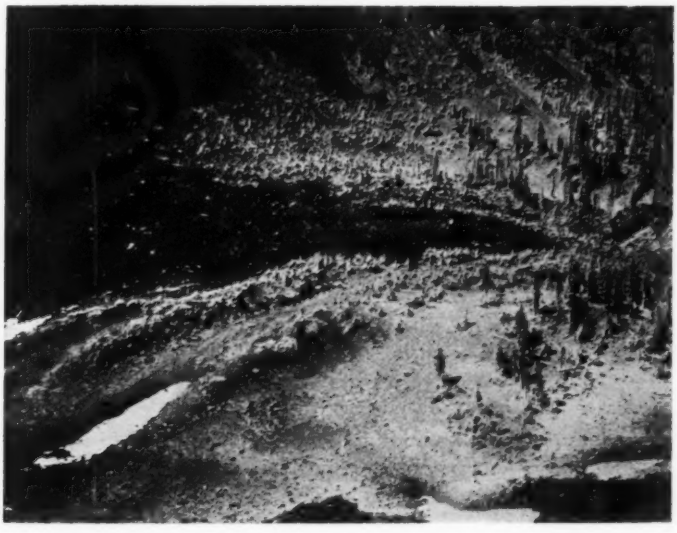

FIG. 6.—Protalus ramparts in cirque at head of north branch of Interior Valley. Note older ramparts near snow and younger ramparts in area of dark rock behind.

was slight and no snowbank occupied the cirque. Most of the ramparts observed elsewhere on San Francisco Mountain belong to the older group.

The age of these protalus ramparts has not been determined. In the glacial chronology of the southern Rocky Mountains the older ramparts would probably be dated as Wisconsin V.[28] However, either set might be related to the period of refrigeration indicated by Matthes'[29] little glaciation or even to some more recent refrigeration.

[28] Ray, *op. cit.*, pp. 1862–63, 1914.

[29] *Op. cit.*

Several seemingly inactive rock streams were observed on the north side of the mountain, one particularly fine example being near the head of Abanaugh Canyon. In this area, at least, such accumulations differ from talus and related deposits only in their more gentle slope and evidence of creeping mass movement. They appear to have been developed during a period of more intense frost action than the present and may correspond to either or both of the colder

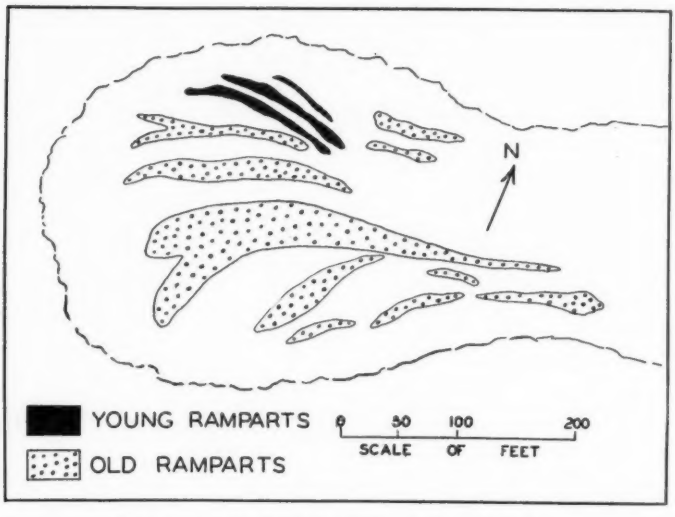

FIG. 7.—Sketch of protalus ramparts shown in Fig. 6

periods indicated by the protalus ramparts. The lower (right-hand) ends of the younger protalus ramparts shown in Figure 6 look as though they might have experienced a slight down-valley creeping movement, probably under the influence of freezing and thawing, suggesting that such accumulations may nourish or even be the source of some rock streams. Robinson[30] mentions a bulky accumulation of rock debris along the north flank of the mountain between 8,700 and 9,000 feet, which he tentatively attributes to solifluction during a former period of refrigeration. A search was made for this

[30] *Op. cit.*, pp. 31–32.

material, but it could not be found, probably because of the heavy forest cover.

Small modern stone stripes similar to those described from New England mountains[31] and elsewhere were observed in many places above timber line (11,500 feet), as were small terracettes, 1–2 feet broad and 3–7 feet long. Both these features are thought to be the product of modern frost action.

Accumulations of pebbles and cobbles brought to the surface by frost heaving cover small soily areas on the floors of cirques. The soil beneath this frost-heaved pavement is 6 inches to 2 feet thick and contains only scattered pebbles and cobbles. Some of the rock fragments in the pavement have come from near-by rock outcrops or large boulders, but these are minor in number and are easily identified by their spall-like form. This frost-heaved pavement is essentially the same as the *Brodelpflaster* described from Europe.[32]

Stone rings, stone nets (*Polygonboden*), and related soil structures[33] are usually attributed to repeated freezing and thawing, although not all workers are agreed on the details of the process. The single example of such features found here is of more than ordinary interest because the date of origin is fixed within narrow limits. An old toll road from Fremont Saddle along the north flanks of Fremont and Agassiz peaks was abandoned in 1932 and is absolutely impassable beyond Fremont Saddle. Crude but distinct stone rings were found on a flat turnout of this road on the flank of Agassiz Peak at 11,600 feet. The material covering the turnout is a loose pebbly rubble of vesicular andesite fragments resting on a 4-inch layer of sandy soil. The best-formed stone ring was 15 inches in

[31] Antevs, *Alpine Zone of Mt. Washington Range* (Auburn, Me.: Merrill & Webber Co., 1932), pp. 53–55; R. P. Goldthwait, "Geology of the Presidential Range," *New Hampshire Acad. Sci. Bull. 1* (1940), pp. 32–40; C. S. Denny, "Stone-Rings on New Hampshire Mountains," *Amer. Jour. Sci.*, Vol. CCXXXVIII (1940), p. 436.

[32] Alfred Dücker, " 'Steinsohle' oder 'Brodelpflaster'?" *Centralblatt f. Min., Geol., Paläo.*, Abt. B (1933), pp. 264–67.

[33] Bertil Högbom, "Über die geologische Bedeutung des Frostes," *Bull. Geol. Inst. Univ. Upsala*, Vol. XII (1914), pp. 308–19; C. S. Elton, "The Nature and Origin of Soil-Polygons in Spitzbergen," *Quart. Jour. Geol. Soc. London*, Vol. LXXXIII (1927), pp. 163–94; J. W. Gregory, "Stone Polygons beside Loch Lomond," *Geog. Jour.*, Vol. LXXVI (1930), pp. 415–18; J Harlen Bretz, "Physiographic Studies in East Greenland," *Amer. Geog. Soc. Spec. Pub. 18* (1935), pp. 175–77.

diameter and more nearly circular than hexagonal in outline. The outer rim of the structure consisted of 2–4-inch cobbles, and the inclosed area was covered by ½–1-inch pebbles. The underlying soil layer was closer to the surface in the central area than under the rim, but the upper surface of the ring showed no noticeable bulging. The structure was not more than 4 inches deep. These stone rings must have been developed within the eight years (1932–40) since the road was abandoned. Small surficial stone rings have been described as forming within a year or two in New Hampshire.[34]

The effectiveness of frost in breaking the relatively brittle and well-jointed volcanic rocks of San Francisco Mountain is shown by the abundance of angular blocks strewn on the surface above timber line and by the filling of cirques and valleys with similar detritus. On the north wall of the Interior Valley, ½ mile upstream from the later Wisconsin terminal moraine, a dike stands fully 75 feet in relief from the surface of the valley wall. This dike could not have so projected during the last glaciation, because it would have been sheared off by the glacier. It must have been left protruding by back-weathering of the less resistant wall rock chiefly under the attack of frost. This gives some measure of the susceptibility of these rocks to frost action and shows why the glacial forms are not better preserved.

WISCONSIN SNOW LINE AND SOUTHERNMOST LIMIT OF GLACIATION IN THE UNITED STATES

The problem of Pleistocene snow lines in western United States has been most thoroughly treated by Klute and Louis.[35] Ray[36] has recently discussed the difference between climatic and orographic snow line, with the natural conclusion that the orographic snow line is the more important in problems of mountain glaciation. The position of the Wisconsin snow line is rather easily determined here, owing to the range in altitude of the peaks composing San Francisco Mountain. Some of the peaks were below snow line and bore no glaciers, while others above snow line were glaciated. Thus, the snow line can be fixed within limits by being approached from two directions.

[34] Denny, *op. cit.*, p. 437.

[35] Klute, *op. cit.*; Louis, *op. cit.*

[36] *Op. cit.*, pp. 1910–11.

The later Wisconsin orographic snow line on San Francisco Mountain was between 11,000 and 11,300 feet. In the Uinta Mountains, of Utah, 350 miles north, the later Wisconsin snow line was close to 10,000 feet.[37] These figures indicate a southward rise in the later Wisconsin snow line of about 300–350 feet in each 100 miles at this longitude (111°40′). Undoubtedly, gradients on snow lines in other areas were different, owing to local influences.

The southernmost extent of mountain glaciation is a matter of some significance because glacial chronology is one of the few ways of dating late geological features in the Southwest. Johnson's[38] brief article on this matter remains the most thorough treatment available. The San Bernardino Mountains of California (34°7′ N. Lat.) have for some time been considered the southernmost area in the United States for which convincing evidence of mountain glaciation has been published.[39] However, Ellis[40] has recently described what appears to be good evidence of glaciation on Cerro Blanco Peak, 12,003 feet high at 33°22′27″, in central New Mexico; and this has been confirmed by Antevs,[41] and H. T. U. Smith and Ray.[42] This is the southernmost point for which seemingly acceptable evidence of glaciation has yet been published. Johnson[43] has properly questioned F. J. H. Merrill's[44] report of glaciation farther south;

[37] Atwood, "Glaciation of the Uinta and Wasatch . . . . ," *op. cit.*, p. 67.

[38] *Op. cit.*

[39] H. W. Fairbanks and E. P. Carey, "Glaciation in the San Bernardino Range, California," *Science*, Vol. XXXI (1910), pp. 32–33; F. E. Vaughan, "Geology of San Bernardino Mountains North of San Gorgonio Pass," *Univ. Calif. Dept. Geol. Sci. Bull.*, Vol. XIII (1922), pp. 335–36.

[40] R. W. Ellis, "Glaciation in New Mexico," *Univ. New Mexico Bull., Geol. Ser.*, Vol. V (1935), pp. 24–25.

[41] Antevs, "The Occurrence of Flints and Extinct Animals in Pluvial Deposits near Clovis, New Mexico, Part II—Age of the Clovis Lake Clays," *Proc. Acad. Nat. Sci. Philadelphia*, Vol. LXXXVII (1935), pp. 306–7.

[42] "Southernmost Glaciated Peak in the United States," *Science*, Vol. XCIII (1941), p. 209.

[43] *Op. cit.*, p. 219.

[44] "Evidences of Glaciation in Southern Arizona and Northern Sonora," *Science*, Vol. XXIV (1906), pp. 116–18.

and G. H. Stone's[45] description of glaciation near Prescott, Arizona, may be dismissed on the basis of later work in that area[46] and present knowledge of snow lines.

San Francisco Mountain (35°20′) is one of the most southerly areas in the United States in which multiple Pleistocene glaciation has yet been reported.

RELATION BETWEEN GLACIATION AND VOLCANIC ERUPTIONS

The volcanic eruptions by which San Francisco Mountain was built ceased early enough in the Pleistocene to permit the cone to be breached by erosion[47] prior to the Illinoian glaciation. If Robinson[48] is correct in assigning these eruptions entirely to the Quaternary, it may be that some phases of this volcanic activity were contemporaneous with pre-Illinoian glaciation. Glaciers may not have formed on San Francisco Mountain, however, either because it was too low or because it was too hot. If such glaciers did form, the evidence has been destroyed or escapes recognition.

The San Francisco Mountain eruptions were followed by a succession of volcanic outbursts which produced the numerous basalt cones and flows in the surrounding area (Fig. 2). Colton[49] divides these post–San Francisco basaltic eruptions into four separate stages on the basis of age, stage 2 being the oldest and stage 5 the youngest, the latter being dated at about A.D. 875 by archeological methods.

The large deposit of Iowan-Wisconsin outwash east of Sugarloaf has been traced to the basaltic area east of U.S. Highway 89, where it passes under a basalt flow belonging to Colton's stage 3. This would date stage 3 and the subsequent eruptions as post-Iowan-Wisconsin and would suggest that stage 2 is pre-Iowan-Wisconsin, although there is no direct proof of this second relation.

[45] "Note on the Extinct Glaciers of Arizona and New Mexico," *Science*, Vol. XIV (1901), p. 798.

[46] T. A. Jaggar and Charles Palache, "Bradshaw Mountains Folio, Arizona," *U.S. Geol. Surv. Atlas* (1905), Folio 126. Also confirmed by personal communication from L. M. Gould.

[47] Johnson, *op. cit.*, p. 219; Robinson, *op. cit.*, p. 41.

[48] *Op. cit.*, p. 93.

[49] *Op. cit.*

SUMMARY

The cirques, moraines, and outwash deposits on San Francisco Mountain are interpreted as the product of three separate glaciations, tentatively dated as later Wisconsin—possibly late Mankato, Iowa-Wisconsin, and Illinoian. The Interior Valley contained a glacier $4\frac{3}{4}$ miles long with a terminus near 8,500 feet during the Illinoian stage, and the glaciers of the following stages were successively shorter. The fivefold subdivision of the Wisconsin recognized in the southern Rocky Mountains does not seem applicable here, although some recessional pauses or slight readvances of the later Wisconsin ice did occur. No incontestable evidence for Matthes' little glaciation was found, but the latest moraines could possibly be attributed to such a post-Wisconsin glaciation.

Frost action has been intense on San Francisco Mountain in the past and is still moderately effective in areas above timber line. The result has been a considerable modification of glacial features and the production of various frost forms. Protalus ramparts of two separate stages are recognized and presumably indicate separate stages of refrigeration, at least one, and perhaps both, of which are post-Wisconsin. A small, surficial stone ring on an abandoned road high on the mountain has been formed within the last eight years by frost action.

The later Wisconsin orographic snow line on San Francisco Mountain was between 11,000 and 11,300 feet. A comparison with data from the Uinta Mountains indicates a southward rise in this snow line of 300–350 feet per 100 miles at this longitude (111°40′). San Francisco Mountain is one of the most southerly areas in the United States in which evidence of multiple Pleistocene glaciation has yet been reported.

The relation between the Iowan-Wisconsin outwash and basaltic flows indicates that a large part of the basaltic flows and cones surrounding San Francisco Mountain are post-Iowan-Wisconsin.

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A STUDY IN APPALACHIAN PHYSIOGRAPHY

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ABSTRACT

George H. Ashley has presented the theory that the present surface in the Appalachians contains no remnant of any of the older peneplains and doubts the existence of more than one peneplain over the whole Appalachian region. The author defends the view of many geologists that remnants of the Schooley (Kittatinny) peneplain still exists on the ridge crests and uplands throughout the Appalachian and New England region.

INTRODUCTION

G. H. Ashley[1] argues that the present surface of the Appalachians contains no remnant of any older peneplains and his hypothesis is supported, he believes, by facts and reasoning included under the following heads: (1) rate of reduction; (2) youthful appearance of the ridge tops; (3) old appearance of the wind gaps; (4) coincidence of present gaps with points of structural weakness; (5) abandoned trails of primitive drainage, now lost; (6) retreat and lowering of escarpments; (7) critical comparisons between the surface and the rock structure of a series of selected regions, revealing how largely the surface reflects rock content and structure and suggesting doubt of more than one peneplain over the whole Appalachian region.

In his discussion of the rate of reduction, Ashley[2] maintains that if the larger anticlines be restored, mountain-making sandstones such as the Pottsville would be raised to 20,000 feet or more above the present surface. If the rocks were folded at the end of Carboniferous time, 200 million years ago, and a peneplain was produced during Triassic and Jurassic time, involving 65 million years, and if the peneplain uplift began at the opening of Cretaceous time (the seaward edge sinking to receive Lower Cretaceous sediments), it would mean that a reduction of 20,000 feet in 65 millions of years took place before uplift, and no appreciable reduction in 135 million years since uplift. To Ashley, this does not seem reasonable. Assuming

[1] "Studies in Appalachian Sculpture," *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), pp. 1401–12.

[2] *Ibid.*, p. 1403.

that the peneplain was not uplifted until Late Miocene time, possibly 8 million years ago, 200 million years would be available for the reduction of 20,000 feet—a reduction rate of 100 feet per million years. At this rate, according to Ashley, there would have been 800 feet of reduction of the uplifted peneplain since Miocene time. This rate applies only to the hardest rocks.

According to Ashley, the Schooley peneplain everywhere in the Appalachians has been reduced at least 800 feet and the widespread erosion surface, so well preserved on the ridge crests and uplands, is not the original peneplain surface but a reflection of it. The same principle applies to the lower surfaces such as the Harrisburg, which he believes must have been lowered scores or hundreds of feet. The weakness of his argument lies in the assumption of a *uniform* rate of reduction of 100 feet per million years. The Appalachian region stood higher than today for enormously long periods of time, and reduction then was doubtless at a faster rate than during the long intervals when the land stood not far above sea-level, late in the Schooley cycle.

George Stose[3] believes that (1) the Schooley peneplain was cut chiefly in Jurassic time; (2) its surface was only slightly uplifted on the Schooley arch in Cretaceous and early Tertiary time; (3) during this time the exposed part of the peneplain was possibly further reduced toward a level plain while fine sands, arkoses, clays, and marls were being deposited on its submerged border; (4) the main uplift of the Schooley arch took place sometime in the mid-Tertiary, and not until this time did effective erosion of the Schooley peneplain surface begin; (5) remnants of this surface still exist on the tops of mountains in the Appalachians because it has been exposed to effective erosion only during later Tertiary and Quaternary time. If, as Stose points out, the Schooley surface was not greatly uplifted and subjected to effective erosion until mid-Tertiary time, the remnants of this peneplain could still exist on the ridge crests and uplands. If the uplift of the Schooley surface during the Harrisburg cycle was gradual and steady (and the irregular vertical distribution in attitude of the wind gaps and water gaps points that way), the streams, be-

[3] "Age of the Schooley Peneplain," *Amer. Jour. Sci.*, Vol. CCXXXVIII (1940), pp. 461–76.

cause of insufficient erosive power, were not able to dissect and remove those portions of the peneplain still preserved on the hard rock.

Ashley[4] points to the youthful appearance of the ridge tops as an argument that the present surface contains no remnants of older peneplains. He states that some of the ridge tops are too narrow for even a good footpath and that such narrow crests are fitted, with their steep slopes on either side, to hasten decay and the removal of the rock. It is true that some of the ridges are narrow crested, but there are many that are broad, flat topped, or rolling. The crest of Kittatinny Mountain is so broad and level in places that one gets the impression of walking on an extensive plain. In places, low, rounded monadnocks appear on it. Abandoned farmhouses on it indicate attempts to cultivate the soil on the gently rolling surface. A walk along the Skyline Trail, from Lehigh Water Gap to Delaware Water Gap, gives one the impression of a peneplain surface. It is difficult to believe that this surface is only a reflection of an older, higher peneplain, worn down hundreds of feet to this lower summit level. It seems almost inevitable that such a flat surface would have been entirely destroyed or reduced to a narrow ridge during the denudation involving 800 feet or more of erosion. Few will deny the existence of remnants of a peneplain, beveling Schooley Mountain, the Cumberland and Pocono plateaus, and the uplands of the New England region. When all the factors that control the rate of erosion are considered, one would not expect the peneplain to be preserved everywhere on the uplands, nor the ridge summits to present everywhere a uniform elevation.

Ashley[5] points to the thin soils and bare rock exposed on the ridge crests and to the little-weathered talus covering the slopes in places. They indicate, he believes, thousands instead of millions of years of weathering of the parent ledges.

In reply, the author calls attention to boulder fields south of the glaciated area,[6] such as the "Devil's Potato Patch," in Little Gap, near Danielsville, Pennsylvania. These boulders are composed of

[4] *Op. cit.*, p. 1404. [5] *Ibid.*

[6] Karl Ver Steeg, "Wind Gaps and Water Gaps of the Northern Appalachians, Their Characteristics and Significance," *Annals N.Y. Acad. Sci.*, Vol. XXXII (1930), pp. 87–220.

hard, resistant, angular Shawangunk conglomerate and lie on the bottom of the notch and are part of a thick accumulation of debris, including soil, which covers the sides as well as the floor. With few exceptions the sides and floors of the wind gaps in the Appalachians are covered with the products of weathering and erosion which have accumulated since piracy took place. Few who have walked over the surface of Schooley Mountain, the Cumberland Plateau, or some of the flat-topped ridges will deny the presence of a considerable amount of soil and weathered material. The absence of thick accumulations of soil on the ridge tops is not necessarily an indication of the youth of the surface. Weathered material could have been removed from the ridge crests to some extent, for the peneplain has slope enough in places for gravity and wash to operate. Furthermore, the Shawangunk conglomerate, a highly siliceous rock, is extremely resistant to the weathering processes and is a very poor soil-maker.[7] The general impression one gets from a study of the beveled surface on the ridge crests of the Appalachians is its extreme old age. Stose[8] comes to similar conclusions: "Accordant level crests of ridges and mountains in the Appalachians represent actual remnants of the Schooley peneplain, little reduced in altitude."

Ashley[9] believes that most of the wind gaps have been deepened considerably, probably hundreds of feet, since they were abandoned. To this, the author replies that because of the great age of the wind gaps, the present altitudes of their floors need not be the original ones. They may have been lowered somewhat by weathering or rarely by headward erosion by streams in the same manner as cols are produced on divides. In the glaciated area they may have been widened and deepened by ice-scour, or glacial deposition may have raised their floors above the original altitude. In some cases the bottoms may be higher than originally, owing to accumulations of talus and other products of weathering which have gradually moved down their sides, thereby covering the original floor with a thick layer of debris.[10] The sides of the wind gaps are usually gently slop-

[7] Polished and grooved ice-scoured surfaces on the Shawangunk conglomerate on Kittatinny Mountain still exist.

[8] *Op. cit.*, p. 476.

[9] *Op. cit*, pp. 1405–6.

[10] Ver Steeg, *op. cit.*, pp. 106–52.

ing and their floors broadly rounded when compared with those of the water gaps. The latter are V-shaped with steeper sides, and bold rock cliffs and ledges are common. The streams in the water gaps are constantly deepening their channels and sweeping away the weathered debris and talus accumulations. Since piracy occurred, the wind gaps have been abandoned to weathering, whereas in the water gaps the slopes are being continually refreshed by erosion. The protective cover of debris on the sides and floors of the wind gaps slows up or prevents further lowering. The well-preserved remnants of the Schooley (Kittatinny) peneplain, extending to the sides of many of the wind gaps, is evidence against the theory that they have been lowered to any great extent. It is difficult to explain how the wind gaps could have been lowered hundreds of feet when the Schooley (Kittatinny) peneplain, everywhere sloping toward the gaps, and below which they have been cut, is still well preserved up to the sides of many of the notches. It seems that if the wind gaps had been deepened and enlarged to any great extent since abandonment, the peneplain on the adjacent ridge crest would have been destroyed or greatly modified.

Ashley[11] points out that some of the notches in the Appalachians, such as Cumberland Gap, Delaware Water Gap, Lehigh Gap, and Schuylkill Gap, are coincident with structural weakness. H. D. Thompson[12] shows that some of the gaps in the Blue Ridge are located where the resistant rocks have relatively narrow outcrops. Others are situated at points where the sinuous crestline of the Blue Ridge is intersected by a thrust fault. It is a fact, nevertheless, that many of the gaps in the Appalachians show no evidence of structural weakness as a controlling factor. Streams cross anticlines and synclines, with apparent disregard of structure.

Ashley maintains that the gaps are not necessarily formed by the downcutting by streams superposed from a peneplain or coastal-plain cover, their courses transverse to the structure of the rocks. He says:

[11] *Op. cit.*, p. 1406.

[12] "Drainage Evolution in the Southern Appalachians," *Bull. Geol. Soc. Amer.*, Vol. L (1939), pp. 1346–47.

The scarcity of large wind gaps, at all comparable in width with the present water gaps, would indicate that the present water gaps had been well established before the mountain crests had been reduced to their present level, again emphasizing the probability of those crests having been reduced by many hundreds of feet.[13]

The writer[14] has pointed out that superposition could not have occurred on the peneplain now represented on the ridge crests of the Appalachians. Johnson,[15] realizing the difficulties involved in superposing the streams in the Appalachians on the Schooley (Kittatinny) surface, has postulated superposition on an extended coastal-plain cover lying on an earlier erosion surface, possibly the "Fall Zone Peneplain,"[16] now entirely destroyed by erosion except where preserved beneath the present coastal-plain sediments. The author would emphasize that the present water gaps are larger than the wind gaps because in the Schooley (Kittatinny) cycle the larger, more powerful streams enlarged their drainage areas by capture of the weaker ones. The master-streams were already large on the Schooley (Kittatinny) surface, and considerable adjustment to structure took place in that cycle. The major streams which established their courses across the ridges in Schooley (Kittatinny) time have continued to enlarge their drainage basins through the Harrisburg and later cycles. The gaps cut in the ridges during the Schooley (Kittatinny) cycle were doubtless largely removed by erosion in the reduction of the ridges to the approximate level of the peneplain. The present notches were cut by streams which occupied broad, open valleys on the Schooley (Kittatinny) surface. This is indicated by the marked slope of the ridge crests toward the major gaps. The evidence seems to support the conclusion that many of the present ridge crests represent remnants of a rolling surface, in the later stage of the cycle of erosion, well preserved over large areas and extending in many instances to the sides of the gaps, which usually meet the ridge crests at sharp angles.

[13] *Op. cit.*, p. 1407.

[14] Ver Steeg, *op. cit.*, p. 218.

[15] Douglas Johnson and Karl Ver Steeg, "Appalachian Studies" (Abst.), *Bull. Geol. Soc. Amer.*, Vol. XL (1928), pp. 131–33.

[16] Henry S. Sharp, "The Fall Zone Peneplain," *Science*, Vol. LXIX (1929), pp. 544–45.

It is difficult to reconstruct the topography of the Appalachian region in the long period which elapsed from the Appalachian revolution to the completion of the Schooley (Kittatinny) peneplain. There could have been readjustment of stream courses to take advantage of structural weakness not now indicated in the present ridges. One must consider the adjustments to structure that took place in the Schooley (Kittatinny) and earlier cycles in order to interpret the origin of the present drainage lines in the Appalachians.

Stose[17] does not discuss the possibility of superposition but concludes that "only one peneplain was formed in post-Paleozoic time in the Appalachians" and that "this peneplain (Schooley) passes beneath the Cretaceous sediments of the Coastal Plain at the Fall Line and is the same as the 'Fall Zone Peneplain.'" This conclusion differs from Shaw, Johnson, Ashley, and Sharp, who have suggested that the Schooley peneplain may be younger than Cretaceous.

Ashley[18] presents evidence to show that the Allegheny front has retreated by erosion and the escarpment has been lowered in the process. The same process, he believes, occurred in the Blue Knob area in Cambria, Blair, and Bedford counties, Pennsylvania. Comparison of the irregular crests of Allegheny Mountain, Negro Mountain, Negro Mountain, Laurel Hill, and Chestnut Ridge, he believes, suggests that any peneplain that once beveled these ridges must have been some distance above them all in order to allow the formation of the present irregularities in the ridge crests. However, it is possible that these mountains were not reduced to the general level of the Schooley peneplain developed on the weaker rock. Evidence is abundant that there were monadnocks which stood well above the general level of the Schooley (Kittatinny) peneplain, a surface far from level but with considerable relief.

It cannot be denied that erosion in the Harrisburg and later cycles has largely destroyed the Schooley (Kittatinny) peneplain and has lowered many of the ridge crests where conditions were favorable for rapid weathering and erosion. However, the Schooley (Kittatinny) surface has not been completely destroyed and there are still large areas where remnants are well preserved. The principles as put forth

[17] *Op. cit.*, p. 461.

[18] *Op. cit.*, pp. 1408–12.

by Ashley for the Schooley (Kittatinny) and older surfaces must likewise be applied to the later peneplains, such as the Harrisburg and Somerville. Furthermore, the latter are developed on weaker rock, such as shale and limestone. It seems inevitable that an erosion surface developed on weak rock would be completely destroyed in the reduction of a region as much as 800 feet. The excellent preservation of broad, flattish remnants of the Harrisburg peneplain over extensive areas in the Ridge and Valley province creates doubt that such a well-preserved surface could be a reflection of a higher peneplain and could have been lowered so much without destroying the existing features so characteristic of peneplains.

STUDIES FOR STUDENTS

DENSITY CURRENTS AS AGENTS FOR TRANSPORTING SEDIMENTS[1]

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ABSTRACT

Density currents are defined, and their origin discussed. Their driving mechanism is identified, and their characteristic behavior explained. "Black blizzards," underflows of turbid water, and *nuées ardentes* are shown to have much in common. Attention is called to a marked difference in the coarseness of sediments transported by atmospheric and aqueous density currents, and some reasons for this difference are suggested. The effect of flocculation upon density currents is discussed briefly. Estimates are given as to the quantity of fine sediment transported by turbid density flows at Lake Mead, by major dust storms, and by the atmosphere following great volcanic eruptions.

INTRODUCTION

During the quarter-century that ended with 1940 the science of weather prediction was reconstructed upon a new foundation—a foundation in which the knowledge of the behavior of density currents was the chief building material. This remarkable rebirth of an established science was the direct result of a temporary breakdown during the first World War of the international machinery for collecting and distributing meteorological data. The Norwegians, under the leadership of V. Bjerknes, began an almost microscopic study of the atmosphere above their own country, as a substitute for the customary world-wide reports.[2] This approach was amazingly successful. The early discovery of the important role played by atmospheric density currents opened the way for these resourceful scientists to develop the "air-mass" system of weather prediction, a method so revolutionary that it met with opposition on almost every hand.

[1] Published with permission of the Soil Conservation Service, U.S. Department of Agriculture.

[2] Sir Napier Shaw, *The Air and Its Ways* (London: Cambridge University Press, 1923), p. 74.

The air-mass concept is truly the master-key that gives access to an understanding of the origin, development, and movement of storms. Air masses in motion are density currents, and the modern meteorologist has made himself an expert on their behavior. Except in his profession, density currents are generally regarded as rare and mysterious occurrences that are of no particular importance beyond a certain academic interest. Outside of meteorological circles, therefore, they continue to be one of the least understood of the more common natural phenomena.

The rapid development and widespread use of eletrical sounding devices have been responsible for the discovery of submarine canyons on the continental shelves in unsuspected profusion. In seeking to penetrate the mystery surrounding their origin, physiographers, oceanographers, and others have advanced nearly a score of hypotheses. Daly's suggestion that they might have been cut by density currents[3] temporarily stimulated geological interest in these flows. Ph. H. Kuenen performed laboratory experiments that he believed supported Daly's hypothesis;[4] but at present the consensus of opinion seems to be that the task is entirely too great for density currents alone to accomplish.

With geological attention thus concentrated upon their apparent impotency in a particular instance, it becomes possible to lose sight of, or to underestimate, their capacity for other work. Experimental research on density currents over a period of years has convinced the writer that they may be of considerable importance as agents of entrainment, transportation, and deposition of the finer sediments, and it is hoped that this paper may refocus geological attention upon them.

No generally acceptable definition of a density current has found its way into the literature; and it seems wise, therefore, to cite a few more or less familiar examples. When a refrigerator door is opened, cold air literally falls out and flows along the floor. Such

[3] R. A. Daly, "Origin of Submarine 'Canyons,'" *Amer. Jour. Sci.*, ser. 5, Vol. XXXI (1936), pp. 401–20.

[4] "Experiments in Connection with Daly's Hypothesis on the Formation of Submarine Canyons," *Leidsche Geologische Mededeelingen*, Vol. VIII, No. 2 (1937), pp. 327–51.

flows may be thought of as miniatures of the atmospheric masses that are called "cold fronts" by meteorologists. The cold Rhine and Rhone rivers enter the comparatively warm waters of Lake Constance and Lake Geneva and continue as subsurface streams to the deepest parts of the lakes.[5] The fresh water of rivers spreads out upon the surface of the sea (Fig. 1), and the salty ocean water may flow many miles inland beneath those same streams. The Rio Grande, when it is made unusually muddy by flash floods from the Puerco and the Salado, may flow for 30 miles or more along the bottom of Elephant Butte Reservoir and pass through the outlet gates without ever losing its identity by mixing with the impounded clear water.[6]

These examples are typical of density flows that occur as natural phenomena. In each instance two like fluid masses are involved. Both are gases, or both are liquids—identical fluids, to be exact, except for slight differences in specific gravity which are attributable to temperature, dissolved matter, or suspended material. Since one fluid mass is heavier than the other, there is nothing extraordinary about the fact that, in response to gravity, it flows beneath the lighter fluid and occupies the lowest available areas.

For the purpose of this paper, and with no thought of presenting a universally acceptable definition, a density current may be described as a gravity flow of a liquid or a gas through, under, or over a fluid of approximately equal density. The phrase "of approximately equal density" is used deliberately, because it possesses a convenient elasticity that is made necessary by the present incompleteness of knowledge in the field under consideration.

By definition, then, a river flowing beneath air is not a density current because it weighs perhaps 800 times as much per unit volume as the overlying fluid. But if that same stream happens to con-

[5] Antonin Smrček, "Experiments on the Motion of Water in the Interior of Large Reservoirs," *Hydraulic Laboratory Practice*, ed. John R. Freeman (New York: Amer. Soc. Mech. Eng., 1929), pp. 510–11; Armin Schoklitsch, *Stauraumverlandung und Kolkabwehr* (Wien: Julius Springer, 1935), p. 52.

[6] L. M. Lawson, "Movement of Silt, Elephant Butte Reservoir," *Reclamation Record*, Vol. X, (1919) p. 411; L. R. Fiock, "Records of Silt Carried by the Rio Grande and Its Accumulation in Elephant Butte Reservoir," *Trans. Amer. Geophys. Union, 1934*, Part II (Washington, D.C.: Nat. Res. Coun. Nat. Acad. Sci., 1934), pp. 468–73.

tinue its flow after entering a lake, as do the Rhine, the Rhone, the Rio Grande, and the Colorado, it may weigh only 1.0008 times as much as the overlying lake water. The densities of the two fluids in this second case are approximately equal, and, again by definition, the submerged stream is a genuine density current.

FIG. 1.—Turbid flood water spreading on the surface of the ocean as three streams enter the Pacific near Del Mar, California, on February 22, 1941. (Photo by Roger Revelle and Manly Natland.)

THE ORIGIN OF DENSITY CURRENTS

Nature has been lavish in providing mechanisms for bringing about changes in temperature, for producing solutions, and for accomplishing suspensions upon the smallest, as well as the grandest, of scales. The tadpole or the insect larva, struggling at the bottom of a pool, may stir up enough mud to cause a density current that persists for minutes only and transports a few centigrams of clay a fraction of a meter. At the opposite extreme a Krakatao may blow off its head and create suspensions that encircle the globe, that have a life measured in years, and that distribute billions of tons of fine volcanic debris. Between these extremes are dust clouds stirred up

by stampeding animals, speeding cars, agricultural implements, and by a thousand other causes; enormous quantities of turbid water produced by great storm waves along shores and over shallow bottoms; "black blizzards" of prairie soil whisked away from farms by winds that scatter it in places where it is unwelcome; and muddy rivers carrying their burdens of sediment into reservoirs and lakes and seas (Figs. 2 and 3).

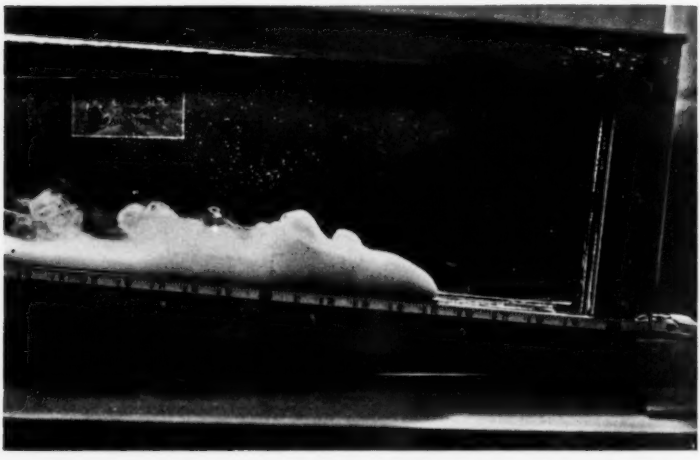


FIG. 2.—This density current of turbid water is a laboratory miniature of flows that transport enormous quantities of fine sediments in reservoirs, lakes, and other bodies of water. The shape of the "head," or "front," is characteristic of density flows.

In such suspensions of fine sediment many density currents have their origin. A current of any kind results from a driving force, and in suspensions this may be supplied in part or entirely by the weight of the suspended particles. Although the contribution of the suspended material to the total force in an ordinary stream is usually insignificant, it may become suddenly dominant under certain circumstances, as, for example, when a muddy river flows into a clear reservoir. In that part of the river channel that lies entirely above the surface of the reservoir, the available driving force may be thought of as the weight per unit volume of the stream, including

the suspended sediment. This does not mean that the entire gravity force is available to overcome frictional resistance to flow, because the river, being immersed in a sea of atmosphere, is buoyed up by a force equal to the weight of the air it displaces.[7] This buoying force is less than 1/800 of the total and is ordinarily ignored.

When the river enters the reservoir, the entire picture changes instantly. Instead of air, the stream now must displace water and is

FIG. 3.—Dust makes a "cold front" visible near Watertown, South Dakota. Dust storms are atmospheric density currents. (Photo by Mrs. M. C. Jorgenson.)

consequently buoyed up by a force equal to the weight of the air plus the weight of the water displaced. For the sake of simplicity, let it be assumed that, upon entering the reservoir, the kinetic energy of the stream is quickly dissipated and that, except for its suspended load, the river water is physically like that in the reservoir. Under the latter assumption the river water, minus its sediment, would be buoyed up by a force equal to its own weight. In other words, the effective weight, or available driving force, of the river

[7] The portion of this available gravity force that is effective in driving the stream is determined by the slope of the channel. In a horizontal channel none of it is effective; in a free waterfall all of it is.

water alone is zero. On the other hand, the suspended sediment does have an effective weight in water, i.e., it weighs more than the water it displaces, and it is this weight that is available as a driving

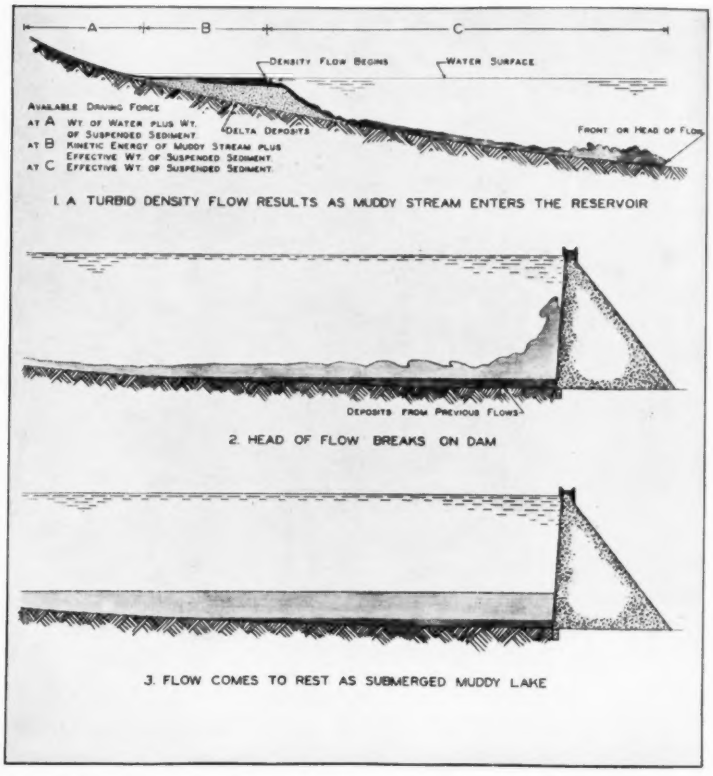

A
B
C
DENSITY FLOW BEGINS
WATER SURFACE
AVAILABLE DRIVING FORCE
AT A WT. OF WATER PLUS WT. OF SUSPENDED SEDIMENT
AT B KINETIC ENERGY OF MUDDY STREAM PLUS EFFECTIVE WT. OF SUSPENDED SEDIMENT
AT C EFFECTIVE WT. OF SUSPENDED SEDIMENT
DELTA DEPOSITS
FRONT OR HEAD OF FLOW
1. A TURBID DENSITY FLOW RESULTS AS MUDDY STREAM ENTERS THE RESERVOIR
DEPOSITS FROM PREVIOUS FLOWS
2. HEAD OF FLOW BREAKS ON DAM
3. FLOW COMES TO REST AS SUBMERGED MUDDY LAKE

FIG. 4.—Diagrammatic representation of the life-history of a density flow

force in turbid density currents. Under the conditions assumed, above, it constitutes the entire driving force (Fig. 4). The reader will find that the terms defined in this paragraph and the one that immediately precedes it are used frequently in the discussions that follow.

In ordinary streams, water propels suspended sediment. In tur-

bid density currents, sediment propels water. Once the significance of this paradox has been grasped, a great deal of the mystery that surrounds density currents is stripped away. It then becomes apparent that the use of a mechanism the very existence of which is, in general, completely ignored, enables these flows to transport and sort the finer sediments. It is for such tasks that they seem to have been designed.[8]

The forces that drive density currents appear insignificant when compared with those found in rivers; yet they are often sufficient to permit the finer silts and clays to deliver themselves promptly to the remotest depths of a lake or reservoir and, under certain circumstances, to deposit themselves far out at sea. In the ocean the journey may involve several or many stops, but in lakes and reservoirs the sediments often travel straight through to their final destination, whether that happens to be 5 miles or 100. Unpublished reports of the United States Bureau of Reclamation provide evidence that many times, in recent years, the turbid waters of the Colorado River have flowed a full hundred miles beneath the clear surface of Lake Mead before coming to rest as a submerged muddy lake held back by Boulder Dam.[9] Those flows did not stop from exhaustion but because they had, literally, no place to go.

The muddy underflow that has as much as 2 per cent of the driving force of the river that created it is a rarity, and, therefore, high velocities are not to be expected. If a driving force as great as 2 per cent does exist, it simply means this: A given volume of the turbid underflow will weigh 2 per cent more than an equal volume of the overlying clear water. When one is inclined to think of this value as being small, it is well to remember that, in the laboratory, density currents continue to flow when the difference in density is a mere 0.01 per cent. If these laboratory flows had taken place under air rather than under water, the driving force would have been 10,000 times as great.

[8] H. C. Stetson and J. Fred Smith, "Behavior of Suspension Currents and Mud Slides on the Continental Slope," *Amer. Jour. Sci.* ser. 5, Vol. XXXV, No. 205 (1938), pp. 1–13.

[9] Nathan C. Grover and Charles S. Howard, "The Passage of Turbid Water through Lake Mead," *Trans. Amer. Soc. Civil Engineers*, Vol. CIII (1938), pp. 720–90.

THE RELATIONSHIP BETWEEN DRIVING FORCE AND VELOCITY

Driving force and velocity, to be sure, do not vary at the same rate. If the depth and other flow conditions remain constant, the velocity varies as the square root of the driving force. Where this relationship holds true, as it does in pipes that are running full, or when two sections of an open channel are identical, except for slope, a 25-fold variation in driving force produces only a 5-fold variation in velocity. Under such conditions the quantity of water carried increases quite obviously, in direct proportion to the velocity.

When a muddy stream is converted into a density flow, the law just expressed does not hold true because the depth no longer remains constant and the quantity of water transported does not vary in direct proportion to the difference in the velocities of the two flows. The comparatively small driving force available in the density current is the primary cause of a great decrease in velocity; and, since the volume of flow is not reduced, the depth must increase. This greater depth tends to make the current flow faster and thus to minimize the effect of the great reduction in the driving force. If it is assumed that there are no other factors that modify the velocity of the density flow, and if it is further assumed that the river channel is rectangular and continues without change along the reservoir floor, it can be shown that the velocity will vary as the cube root of the driving force. A river having a driving force 125 times greater than the density current it produced would have, under these assumptions, a velocity only 5 times as great; and a 10,000-fold difference in driving force would produce approximately a 21-fold variation in velocity.

Actually, the problem is much more complex than these assumptions have made it; and the task of solving it may, possibly, be a pleasure for those who are experts in fluid mechanics. The depth of the density current is increased not only by the change in velocity but also by an increase in volume that results from mixing clear water with the muddy river as it enters the reservoir. A further increase in depth may be attributed to the fact that the overlying water provides a boundary surface at which the resistance to flow is greater than it would be at the same velocity if air were the over-

lying fluid. Additional factors are known to be involved; and, although exact quantitative evaluation is not yet possible, the available data indicate that density-current velocities are somewhat lower than those obtained by use of the cube-root relationship.

Certain measurements from the Colorado River, when combined with a few estimates, allow rather crude but interesting calculations to be made as to the mean velocity of a particular density current of October, 1935, that passed through Lake Mead when storage had been in progress for eight months and the lake was approximately 80 miles in length.[10]

A brief history of that density flow can be reconstructed from the records of three gauging stations. On September 29 the sediment concentration at Bright Angel, 265 miles above Boulder Dam, showed a great increase over that of the previous day, although the discharge had fallen from 13,000 cubic feet per second to 9,630.[11] Eight days later, on October 6, the water became muddy at Willowbeach, 10 miles below Boulder, after having been clear for the previous three weeks. At the time, 9,900 cubic feet per second were being released from the reservoir. On October 8 there was a marked increase in the suspended sediment at Topock, 105 miles below Willowbeach.

If the records indicated at what time the muddy flow reached the reservoir and the dam, the mean velocity of the density current could be determined immediately. Lacking this information, estimates must be made from such data as are available.

According to the records, it required approximately 192 hours for the muddy water to flow from Bright Angel to Willowbeach—a distance of 275 miles. This journey may be divided into three stages: 185 miles of normal channel flow from Bright Angel to Lake Mead, 80 miles of density flow through the lake, and 10 miles of normal flow from Boulder Dam to Willowbeach.

If the time required for the flow to pass through any two of these sections can be determined and subtracted from the known total elapsed time, the remainder represents the value for the third sec-

[10] *Ibid.*, p. 725.

[11] "Surface Water Supply of the United States, 1935, Part 9: Colorado River Basin," *U.S. Geol. Surv. Water-Supply Paper 789* (1937), p. 21.

tion. Since the laws that govern open-channel flow are reasonably well established, while those pertaining to density flows are inadequately understood, the best chance to solve the problem lies in the determination of the time needed to pass through the two open-channel reaches.

An examination of the data shows that the flood traveled from Willowbeach to Topock at a mean velocity of 3.2 feet per second. It may be assumed that a similar velocity prevailed for the 10 miles from Willowbeach upstream to the dam and that, consequently, about 4.6 hours were required for this part of the journey. The data also show that the quantity of water flowing past Bright Angel and Willowbeach were essentially the same and that the mean slope of the channel from Bright Angel to Lake Mead was 5 times as great as from Boulder Dam to Topock. Taking into consideration this difference in slope and the probable difference in channel width, it is reasonable to assume that a given flow would move about twice as fast above the reservoir as it would below the dam. If, then, a velocity of 6 feet per second is assumed for the reach above the reservoir, approximately 45.2 hours would be required for the stream to flow from Bright Angel to Lake Mead. After subtracting the time consumed in both open-channel sections from the total elapsed time, there remain 142.2 hours for 80 miles of density flow within the reservoir. This indicates a mean velocity of 0.83 feet per second. Certainly this is a reasonable value.

When the problem is attacked upon the basis of the effective driving force, sediment-concentration data are of primary importance. On October 6 the Willowbeach samples contained 1.57 per cent sediment by weight. Eight days earlier the concentration was 2.14 per cent by weight at Bright Angel; and a cubic foot of river water at that station weighed 63.24 pounds, which is 103 times as great as the effective weight of the sediment contained in an equal volume of river water at Willowbeach on October 6. If it is assumed that the concentration at Willowbeach was also the actual mean concentration in the density current, the true ratio of the driving forces in the river and the density current will be approximated when 103 is multiplied by the ratio of the mean slopes of the river and the reservoir, which, in this instance, is 2.22 and makes the corrected

value nearly 230. From the cube-root relationship the ratio of the velocity of the river to that of the density current is about 6.12, the cube root of 230. Knowing the length of Lake Mead, the distance from it to Bright Angel, and the time required for the silty water to flow from Bright Angel to Willowbeach, it now may be determined that the mean velocity of the density current was 0.86 feet per second. This value checks so closely with the one previously obtained (0.83) that it becomes important to remember that it is based upon a series of assumptions and, therefore, should not be considered as unquestionable proof of the dependability of the analysis.

At Lake Arthur Reservoir in South Africa[12] and at Elephant Butte Reservoir[13] observers have estimated that currents of turbid water move along the reservoir floors at 0.5 feet per second or slightly less, and a similar flow is reported to have passed through Echo Reservoir, Utah, at a velocity of 0.22 feet per second.[14] As far as the writer knows, the literature contains no precise data on this subject, and this merely serves to emphasize the fact that the task of obtaining such field information is a difficult one. The values given above are means, and, although they may seem too low to be effective, they have been competent to carry fine sediments completely through the various reservoirs, regardless of length. Apparently nothing has been published regarding maximum velocities; but, judging by laboratory experiments in which density currents were observed in channels having both broad sections and narrow ones, the maxima may be 3 or 4 times as great as the means. Even so, they can scarcely be thought of as high velocities.[15]

[12] A. D. Lewis, discussion of the Grover and Howard paper cited above, *Trans. Amer. Soc. Civil Engineers*, Vol. CIII (1938), pp. 757–59.

[13] John H. Bliss, discussion of Grover and Howard paper, *op. cit.*, p. 650.

[14] R. E. Redden, discussion of Grover and Howard paper, *op. cit.*, p. 772.

[15] The writer's experiments have been conducted, for the most part, in a glass-walled flume 5 feet long, 10 inches deep, and 2.25 inches wide. Two larger flumes have been used occasionally: one is 10 feet long and 10 inches square; the other is 15 feet long, 34 inches deep, and 15 inches wide. Underflows having effective densities (weight per unit volume of underflowing liquid divided by weight per unit volume of overlying liquid) from 1.0001 to 1.0080 have been investigated. Bottom slopes of from 0.25 to 8.0 per cent have been used; and temperatures, carefully controlled, have varied from

Slowness of motion, however, is not an inherent characteristic of density currents. Masses of extremely cold air accumulate over the plateaus of Greenland and Antarctica and flow seaward as "gravity winds" at speeds that often exceed 100 miles per hour.[16] The muddy underflows of lakes and reservoirs have their aerodynamic counterparts in dust storms, and the largest of these involve hundreds of

FIG. 5.—A "black blizzard" sweeping into Spearman, Texas, April 14, 1935. (Photo by F. W. Brandt.)

thousands or even millions of cubic miles of atmosphere[17] in which velocities of from 30 to 60 miles per hour (44–88 feet per second) are often encountered (Figs. 5 and 6). According to press reports, a black blizzard was accompanied by 85-mile winds (124 feet per second) in the Texas Panhandle during February, 1941. This velocity

45° to 95° F. Mean velocities of the currents only rarely have exceeded 4 inches per second. Density differences ordinarily have been obtained by the use of fine sediments in suspension, although sugar, sodium chloride, and sodium thiosulphate have been used occasionally, as have differences in temperature.

[16] George F. Taylor, *Aeronautical Meteorology* (New York: Pitman Pub. Co., 1938), p. 93.

[17] C. B. Blacktin, *Dust* (Cleveland: Sherwood Press, 1934), p. 39.

is approximately 20 miles greater than any previously reported for dust storms in that region.

In December, 1929, one of the hot clouds (*nuées ardentes*) that occasionally rush down mountain sides following volcanic explosions

FIG. 6.—This dust storm was about 2,000 feet deep as it rolled across Nebraska. Depths of over 20,000 feet have been recorded. (Photo by Captain C. E. Recknaegel.)

was observed at Mont Pelée as it traveled a 4-mile course at a mean velocity of 96 miles per hour (140 feet per second).[18] These clouds ordinarily move much more slowly[19] and are apparently true density flows in which suspended particles furnish the driving force (Fig. 7).

[18] Frank A. Perret, "The Eruption of Mt. Pelée, 1929–1932," *Carnegie Inst. Wash. Pub. No. 458* (1935), p. 19.

[19] Velocities less than 25 miles per hour are noted by several observers, including Perret and Heilprin.

FIG. 7.—A *nuée ardente* at Mount Pelée. Such "fiery clouds," composed of volcanic dust and hot gases, have much in common with dust storms and turbid underflows. (Photo by Frank A. Perret, courtesy Carnegie Institution of Washington, D.C.)

Edmund Otis Hovey holds a somewhat different opinion and states that the "exploding clouds of steam were so overloaded with dust and larger fragments of comminuted lava, that they flowed down the slope of the mountain and the gorge, like a fluid propelled at a high velocity by the horizontal or partly downward component of the force of the explosion."[20] Perret describes the clouds as "a mass of discrete particles separated by vapor films, giving the aggregate a mobility quite frictionless," and again states that such a mass flows "down any slope with incredible speed because it is frictionless; the extreme manifestation of power in explosive action." He observed, too, that thousands of explosions took place within some flows. T. Anderson and J. S. Flett, in writing about the Soufrière eruption of 1902, record that a huge black cloud rolled down the mountain in globular surging masses, moving like a current of heavy gas.[21]

All are describing concentrated suspensions of particles flowing under density-current conditions. According to R. H. Finch, considerable volumes of hot, gas-charged dust, sand, and pumice are a "requisite characteristic of all *nuées ardentes*,"[22] and C. N. Fenner states that gravity would act upon such a mixture just as it would upon a liquid, and cause it to flow.[23] The primary task of the explosions is, seemingly, to create suspensions rather than to propel them after they have been forced beyond the rim of the crater; and, whether or not this is the case, the fact remains that the suspended particles constitute a driving force capable of moving such "fiery clouds" at either low or high velocities without other assistance. As a matter of fact, Perret has observed that rain on the mountain, by supplying torrents of water to dry channels filled with hot sand and volcanic ash, occasionally produces clouds of steam in which large

[20] "New Cone and Obelisk of Mont Pelé," *Bull. Geol. Soc. Amer.*, Vol. XV (1904), pp. 558–60.

[21] "Report of the Eruption of the Soufrière in St. Vincent in 1902," *Phil. Trans. Roy. Soc. London*, Ser. A, Vol. CC (1903), p. 394.

[22] "On the Mechanics of *Nuées Ardentes*," *Jour. Geol.*, Vol. XLIII, No. 5 (1935), pp. 545–50.

[23] "The Origin and Mode of Emplacement of the Great Tuff Deposit of the Valley of Ten Thousand Smokes," *Nat. Geog. Soc. Katmai Ser.*, No. 1 (1923), p. 72.

quantities of fine particles are thrown into suspension.[24] These clouds flow down the slopes and proceed for several miles out over the ocean unaided by explosions.

THE BEHAVIOR OF DENSITY CURRENTS

It is easy to lose sight of the fact that turbid underflows and dust storms are manifestations of a single phenomenon and that, therefore, they may be expected to be alike in behavior. A muddy density current is, essentially, nothing more or less than a submerged dust storm which, in turn, is merely a turbid underflow in a sea of atmosphere. A dust storm, like any wind, leaps lightly over barriers that would effectively check a river, and almost completely ignores meanders that would virtually enslave a stream. Such behavior is commonly observed and, therefore, seems perfectly natural.

Winds—clean or dirty, moist or dry, hot or cold, violent or gentle—are density currents and behave as density currents. But turbid underflows are density currents also and, consequently, they do not behave like muddy rivers but like dusty winds. If this behavior seems unnatural, it is because we have failed to comprehend that the governing factors are not the great differences that exist between masses of air and masses of water, but the comparatively minute ones that exist between two masses of air or two masses of water. The understanding of liquid underflows is sometimes difficult because, unwittingly, we have granted to gases—particularly to the atmosphere—the exclusive right to behave as density currents. Nature recognizes no such division and, consequently, we may find ourselves a bit puzzled when a liquid mass seems to insist upon behaving as if it were a gas.

If it were possible to vary the slope of a river channel so that a given volume could be made to flow through it under air and under water at the same velocity, the inertia forces would be equal in the two flows. These forces might be great enough to make underflows leave the channel at curves and leap over obstacles and yet be comparatively impotent in open streams at the same velocity. In the latter case they are opposed by normal gravity forces, whereas in density flows the effectiveness of gravity is largely nullified by the

[24] Perret, *op. cit.*, pp. 28, 53, 98, 116.

buoying effect of the displaced reservoir water. Evidently, then, the behavior of each flow is governed very largely by the ratio of its gravity and inertia forces. The ratios for a muddy river and a turbid underflow are very different, and so are their behaviors; but the ratio for a wind is of the same order of magnitude as that for a turbid underflow, and it would be strange if there were not marked similarities in their behavior. No clearer impression can be given of the behavior of a muddy density current passing through a reservoir than is provided by comparing it to a slow-motion picture that might have been made of a shallow, dusty breeze moving down the same valley before the dam was built and a lake impounded. Unless the channel is very deep, the flow is apt to abandon it at even moderately abrupt bends; a seemingly impassable transverse barrier, such as a coffer dam, may be cleared in an apparently effortless leap (Fig. 8, *B*); and the turbid water will rush through channel restrictions to spread out again and move more deliberately in wider reaches.

In recent years much time and attention have been devoted to considering the possibility that density flows may destroy themselves by mixing with the overlying fluid.[25] The boundary between the two fluids commonly is so distinct that it forms a definite interface upon which waves are more or less abundant.[26] More than half a century ago H. Helmholtz pointed out that if the difference in the velocities of the two fluids is greater than the velocity of these waves it becomes possible for mixing to occur at the interface.[27] This is apparently verified by laboratory studies which show that, as the difference in the velocities is increased, the interfacial waves grow shorter and less rounded, until they become cusped and, if they are dragged downstream, their crests are swept off by the overlying clear water, in much the same manner that a high wind sweeps the crests from ordinary surface waves.

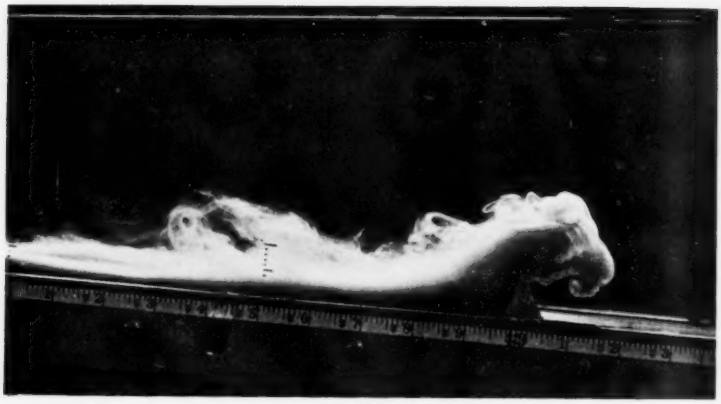
This may lead to the conclusion that when such mixing is present

[25] B. H. Monish, discussion of the Grover and Howard paper, *Trans. Amer. Soc. Civil Engineers*, Vol. CIII (1938), pp. 751–55.

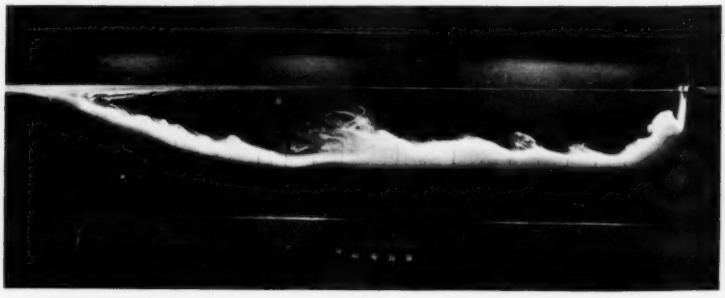
[26] When one looks down upon a layer of cloud from a mountain height or an airplane, he sees exactly such an interface.

[27] *Wissenschaftliche Abhandlungen*, Vol. I (Leipzig, 1882–83), p. 146.

A



B



C

FIG. 8.—(*A*) A muddy stream enters the reservoir and is changed to a density flow which moves along the delta. (*B*) The underflow leaps a channel obstruction. (*C*) The turbid water climbs toward the surface as it strikes the dam.

a density current rapidly destroys itself through dilution. This, in turn, implies that dilution makes the driving force ineffective. In turbid underflows this force may be supplied solely by the effective weight of the suspended sediment, and, therefore, any change that causes deposition tends to destroy flows of this kind. But, certainly, the first effect of dilution appears to be in the opposite direction, since the depth of the turbid water is actually increased, and thus a portion of the sediment is suspended at elevations higher than those it formerly occupied.

It is recognized generally that the ability to keep sediment in suspension is intimately related to the velocity of the transporting stream, and it should be kept in mind that this relationship is extremely close in those turbid underflows that owe whatever velocity they attain to the presence of the suspended particles themselves. If dilution reduces the velocity of flow, deposition may take place; and if the time available for such deposition is long enough, the current will be destroyed. However, the settling velocity of the particles involved is so slow that the flow may reach its destination long before complete deposition is accomplished.

Kuenen has stated that for a given bottom area it makes no difference whether a given quantity of sediment is suspended in a layer of water 1 meter or 100 meters thick: the resulting density currents will flow at the same velocity.[28] Apparently, he assumes that the channel is of very great width in terms of the depth of flow and that, in each case, the sediment distribution is homogeneous. Under those conditions the effect of any increase in depth produced by dilution will be exactly balanced by the accompanying decrease in the effective density of the turbid water. When the channel is not many times the depth of the density flow, the velocity may be expected to decrease because the increased resistance offered by the sides is no longer negligible.

The problem is complicated immensely by possible changes in the effective settling velocities of the suspended particles that dilution might produce by altering the degree of flocculation, by the nature of the velocity distribution and the sediment distribution within the

[28] *Op. cit.*, pp. 346–47.

flow, and by a multitude of other factors. In nature the velocity will certainly be influenced by changes in the slope, width, and roughness of the channel and probably by differences in temperature and salinity in various reaches of the lake or reservoir. Regardless of the influence of all these factors, many turbid underflows, which must have experienced dilution from continuous or intermittent interfacial mixing for 100 miles or more, have not been destroyed in Lake Mead before reaching Boulder Dam.

Laboratory observations indicate that, although such dilution may increase the depth of turbid water many fold, it does not appear to change the depth of the original underflow. The elevation of the very active interface remains virtually unchanged, because the mixing process adds clear water to the underflow at essentially the same rate that it takes silty water from it. As the water above the interface becomes increasingly turbid, its newly acquired sediment load drives it slowly along as a density current superimposed upon, and dragged along by, the original underflow. Although these two flows are in contact, they are quite distinct and are separated by the interface. As the mixing process continues, the mean sediment concentrations above and below the interface approach the same value, and so also do the mean velocities in the two flows. It seems probable that these two factors combine to reduce the rate of dilution and thus to lengthen the time required for the two flows to unite and behave as a single density current. In the writer's opinion field observations will show that such a union is rarely accomplished before the underflow reaches its destination and brings the mixing process to an end; but, certainly, a prolonged inflow of water, or special channel conditions such as a submerged fall, would greatly increase the probability of a complete union of the two flows.

In many reservoirs where underflows occur, the surface waters remain clear regardless of mixing at the plane of contact between the two flows, so that the original interface ceases to be the dividing-line between clear and turbid water and becomes merely the line of demarcation between two masses of water that differ in the degree of their turbidity. A new interface thus is developed between the upper turbid layer and the overlying clear water. It is this comparatively inactive interface that the field observer sees; and upon it

waves are ordinarily very long and shallow if present, but they may be entirely lacking. Velocity differences at this level are apt to be small and thus tend to give an entirely false impression of conditions within the flow.

A surface phenomenon of particular interest frequently accompanies turbid underflows. Floating debris accumulates near the head

FIG. 9.—An "island" of driftwood at the beginning of underflow in Lake Mead. (Photo by U.S. Bureau of Reclamation.)

of the reservoir immediately downstream from the place at which the inflowing muddy water seems to plunge beneath the clear surface (Figs. 9 and 10). This locality is called *Brech* at Lake Constance and *La Bataillère* at Lake Geneva. It marks the point at which the kinetic energy of the inflowing stream is no longer able to drive the muddy water into the reservoir and marks, also, the beginning of the underflow. As the two masses of water come into contact, a part of the stream's kinetic energy is used to mix considerable quan-

tities of clear reservoir water into the turbid inflow. This water is carried away by the density current, along with other clear water that is dragged along immediately above the interface. To replace it, a pronounced upstream current develops at the reservoir surface and prevents driftwood and other floating material from dispersing as it normally would. The reservoir apparently refuses to accept the debris; and, since the river won't take it back, it is impounded

FIG. 10.—A sharp line of demarcation between the muddy water of the Colorado River and the clear water of Lake Mead at the beginning of underflow. (Photo by U.S. Bureau of Reclamation.)

just below *La Bataillère* for the duration of the density flow. Such a floating island is a characteristic symptom accompanying violent underflows.

THE SIZE OF TRANSPORTED PARTICLES

A small amount of quantitative data on the size of particles transported by density currents is available. A single large sample taken at the outlets of South Africa's Lake Arthur Reservoir during an underflow in November, 1935, contained "a few stray grains of sand," but 72 per cent of the suspended sediment was finer than 5 μ, and 60 per cent was finer than 2 μ.

It is of considerable interest, then, that in a series of twenty-six

samples taken from density-current deposits in lower Lake Mead 77 per cent of the sediment was finer than 5 μ and that 97.7 per cent was finer than 20 μ (Table 1, sample 12). Particles as large as 50 μ were entirely lacking. These samples were obtained by the United States Bureau of Reclamation, midway between the intake

TABLE 1

PARTICLE SIZE DISTRIBUTION FOR SAMPLES OF DUSTFALLS LOESS, AND UNDERFLOW DEPOSITS

Particle Diameter for Dustfalls and Loess	Size Distribution in Per Cent by Weight*												Particle Diameter for Underflow Deposits
	Dustfalls				Loess				Underflow Deposits				
	1	2	3	4	5	6	7	8	9	10	11	12	
$<5\ \mu$	33.4	25.57	11.15	24.19	32.0	24.5	10.5	23.0	81.3	76.9	70.2	77.0	$<5\ \mu$
$5-10\ \mu$	31.7	11.91	22.01										
$10-25\ \mu$	32.1	44.09	56.17	72.60	56.4	66.5	42.3	....	16.9	21.0	26.8	20.7	$5-20\ \mu$
$25-50\ \mu$	0.6	11.35	5.99										
$>50\ \mu$	2.6	5.99	2.99	3.21	11.5	9.0	47.2	....	1.7	2.1	3.0	2.3	$20-50\ \mu$

* Explanation of samples:

1. Dust from brown snow, New Hampshire and Vermont, February 24, 1926. W. O. Robinson.
2. Dustfall at Madison, Wisconsin, March 19, 1920. A. N. Winchell and E. R. Miller.
3. Dustfall at Madison, Wisconsin, March 9, 1918. A. N. Winchell and E. R. Miller.
4. Average for samples from five American dustfalls, including the three above.
5. Loess, 3 feet below surface, Grundy County, Missouri. *Field Operations, U.S. Bur. Chem. and Soils, 1914*, p. 1991.
6. Loess, 3 feet below surface, Muscatine County, Iowa. *Field Operations, U.S. Bur. Chem. and Soils, 1914*, p. 1848.
7. Loess, 3 feet below surface, North Platte, Nebraska. W. W. Barr, *Neb. Agr. Exp. Sta. Res. Bull. 5* (1911), p. 12.
8. Average for ten loess samples from Iowa, Missouri, Nebraska, Mississippi.
9. Bottom deposit, Boulder Canyon, Lake Mead, March 23, 1939, U.S. Bureau of Reclamation.
10. Bottom deposit, Boulder Canyon, Lake Mead, June 23, 1939. U.S. Bureau of Reclamation.
11. Bottom deposit, Boulder Canyon, Lake Mead, December 7, 1939. U.S. Bureau of Reclamation.
12. Average for twenty-six samples laid down by density flows at Lake Mead. U.S. Bureau of Reclamation.

towers at Boulder Dam, during a period of eighteen months. They were remarkably alike. In none was less than 70.2 per cent or more than 81.3 per cent of the sediment finer than 5 μ (Table 1, samples 9 and 11). One sample had no particles coarser than 20 μ, but in the remaining twenty-five samples none had less than 1.2 per cent or more than 3.9 per cent coarser than 20 μ.

Lake Arthur is 6 miles long and has a mean bottom slope of 15 feet per mile. Lake Mead is nearly 120 miles long and has a mean

bottom slope of approximately 5 feet per mile. It is entirely possible that the marked similarity in the data is little more than a coincidence. In oceans, lakes, and even in small reservoirs, bottom slopes of several hundred feet per mile are to be found. On such slopes there is every reason to believe that much coarser sediments are readily transported by aqeuous density currents.

Only a very few comparable mechanical analyses are available for deposits laid down by recent dust storms.[29] If the division is again made at 5 μ, the values are found to be almost exactly reversed, since the five available analyses show that 24.2 per cent of the sediment is finer than 5 μ, as compared with 72 and 77 per cent at Lake Arthur and Lake Mead. As might be expected, variations among the samples are much greater than for water-transported sediments; a dustfall sample collected from snow at Madison, Wisconsin, contains only 11.15 per cent finer than 5 μ, but one from snow in the New England states shows almost exactly 3 times as great a concentration—33.4 per cent (Table 1, samples 1 and 3).

It is of interest that analyses of ten samples taken from loess deposits of the Mississippi Valley give nearly identical figures both for the mean percentage of particles smaller than 5 μ (23.0 per cent) and for the extremes (10.5 and 32.0 per cent) (Table 1, samples 5, 7, and 8).[30]

In 1898 J. A. Udden published a mechanical analysis for a deposit of volcanic dust that fell on snow in Norway following an eruption in Iceland.[31] Although the distance from Iceland to Norway is over 600 miles, more than 99 per cent of the particles were coarser than 16 μ; 10 per cent were coarser than 125 μ; there were practically none larger than 250 μ.

That sediments laid down by eolian density currents should be

[29] W. O. Robinson, "Composition and Origin of Dust in the Fall of Brown Snow, New Hampshire and Vermont, February 24, 1936," *Monthly Weather Review*, Vol. LXIV, No. 3 (1936), p. 86; A. N. Winchell and Eric R. Miller, "The Dustfall of March 9, 1918," *Amer. Jour. Sci.*, ser. 4, Vol. XLVI (1918), pp. 599–609, and "The Great Dustfall of March 19, 1920," *ibid.*, ser. 5, Vol. III (1922), pp. 349–64.

[30] F. Leicester Cuthbert, "Petrography of Two Iowa Loess Materials," *Amer. Min.*, Vol. XXV (1940), pp. 519–27.

[31] "The Mechanical Composition of Wind Deposits," *Augustana Libr. Pub. I* (1898), p. 36. Quoted in Winchell and Miller, "The Great Dustfall . . . . ," *op. cit.*, p. 362.

coarser than those deposited by similar aqueous flows is somewhat of a surprise to many interested persons. Even a casual observer knows that normal streams can and do move very coarse materials, and that the most abnormal winds rarely transport pebbles much larger than peas. Such common knowledge is probably largely responsible for this apparently mistaken idea as to the relative coarseness of the deposits laid down by atmospheric and by liquid density currents.

Of the many factors that influence the character of such deposits, the more important are (1) the relative settling velocity of the sediments in air and in water, (2) the relative suspending power[32] of the transporting fluids, and (3) the conditions under which deposition takes place.

From Figure 11 it can be seen that particles with diameters of less than 20 μ settle about 90 times as fast at 20° C. in air as in water, and that the lower the temperature the higher this ratio becomes. From this relationship it would appear that water should transport the coarser material, and so it would if other things were equal.

The suspending power of a current is influenced chiefly by velocity, density, and viscosity. Of these three factors, the first would appear to be by far the most important, for, although water, as we find it in nature, is somewhat more than 800 times as dense as air and has an absolute viscosity from 30 to 120 times as great, it seems incapable of keeping any but the very finest of sediments in suspension in density currents. In atmospheric flows, particularly in dust storms, the mean velocities may be from 30 to 60 or more times as great as in turbid underflows, and, as the evidence shows, the suspending powers they create are great enough to entrain and transport relatively coarse sediments for vast distances.

There is another modifying influence that may be of even greater importance. A sample of a bottom deposit from a lake or reservoir probably supplies a rather complete and trustworthy account of the sediment load of a turbid underflow, but a sample of a dustfall or

[32] Assuming that sediments of all sizes are available for transportation by density currents, "suspending power" may be thought of in terms of the coarsest inorganic material that is carried in appreciable quantities—say 1 per cent or more by weight.

of loess quite certainly tells an incomplete and biased story about the load carried by the atmosphere. The reason is rather simple but, perhaps, not too obvious. The underflow ordinarily comes to rest in a basin and deposits its entire load there, whereas the dust storm merely flows over a basin into which a part of its load settles. In one case the load and the deposit are identical, and a wisely taken sample

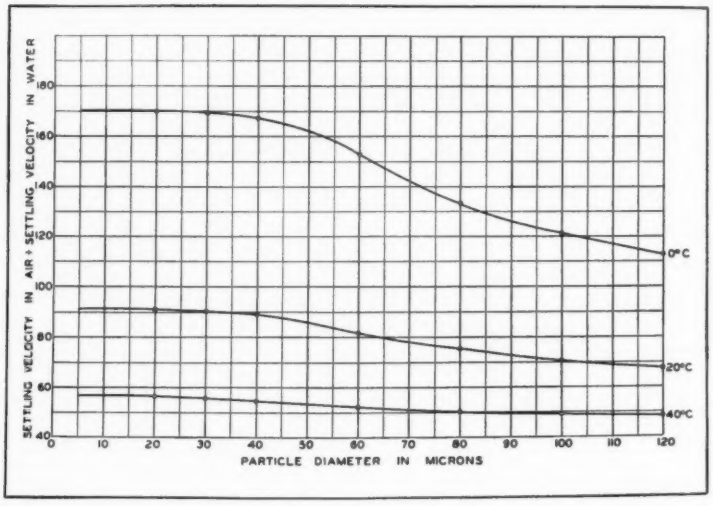


FIG. 11.—The relative settling velocity of quartz spheres in air and in water plotted against particle diameter. Settling-velocity data obtained by applying Stokes's law and from laboratory determinations as presented in *Handbuch der Experimentalphysik*, IV, No. 2 (Leipzig: Akademische Verlagsgesellschaft M.B.H., 1932), pp. 367 and 369.

is adequate for either; but in the other case, no matter how skilfully a sample may be taken, it can tell only what was deposited, not what was being transported. Coarse particles settle out of the atmosphere promptly, but those that are a micron or less in diameter may be carried about for months or even years.[33]

The extreme coarseness of volcanic dust deposits hundreds of miles from their sources, as indicated by Udden's mechanical analy-

[33] Theodor von Kármán, "Turbulence," *Jour. Roy. Aero. Soc.* (December, 1937), pp. 1–34. See esp. p. 12.

sis cited above, is due in large measure to the method of entrainment. In violent eruptions particles are blasted to enormous heights. In contrast to American dust storms, which rarely carry dust to heights of 20,000 feet, Vesuvius produced an ash cloud in 1631 that, according to measurements by Braccini, reached a height of 48 kilometers (29.8 miles).[34] A particle 100 μ in diameter would require nearly ten days to settle from so great a height and consequently could travel to some very distant point before it was deposited.

Sight should not be lost of the facts that data on the sizes of particles in sediments laid down by density flows are meager and that, in the case of turbid underflows, the range of field conditions thus far studied is exceedingly limited. Bottom slopes in lakes and reservoirs are ordinarily not more than a few feet per mile, and such data as are now available come from just such places. As already stated, slopes in oceans, lakes, and small reservoirs occasionally must be measured in hundreds of feet per mile, and velocities attained by density flows occurring under such conditions very well may be several to many times as great as any known to prevail in reservoirs. It would not be unreasonable, therefore, to expect greatly increased suspending power in certain density flows and, consequently, a very marked increase in the coarseness of the transported sediments.

THE EFFECT OF FLOCCULATION

The extreme fineness of the sediments transported by turbid underflows very naturally raises the question as to the effect of flocculation. There is apparently little field evidence to support the belief that flocculation destroys density flows that are caused by suspended sediment. Actually, at the three great reservoirs from which most of the available information has been obtained, the material in suspension seems invariably to be more or less highly flocculated. Lewis, who has supplied the data from Lake Arthur Reservoir, inclines to the belief that flows of this sort are "essentially associated with flocculated clay."[35] On the other hand, the writer's laboratory experiments have been carried out without difficulty, although suspensions of completely dispersed clay particles ordinarily were used.

[34] W. J. Humphreys, *Physics of the Air* (New York: McGraw-Hill Book Co., 1929), p. 596.

[35] Lewis, *op. cit.*, p. 757.

The suggestion has been made that turbid underflows are not possible when a muddy river flows into the ocean, because the less dense muddy water may spread out upon the surface of the salt water and then settle to the bottom in a highly flocculated condition. Under such conditions it has been observed that the muddy water breaks through the interface and settles in small masses, just as the warm

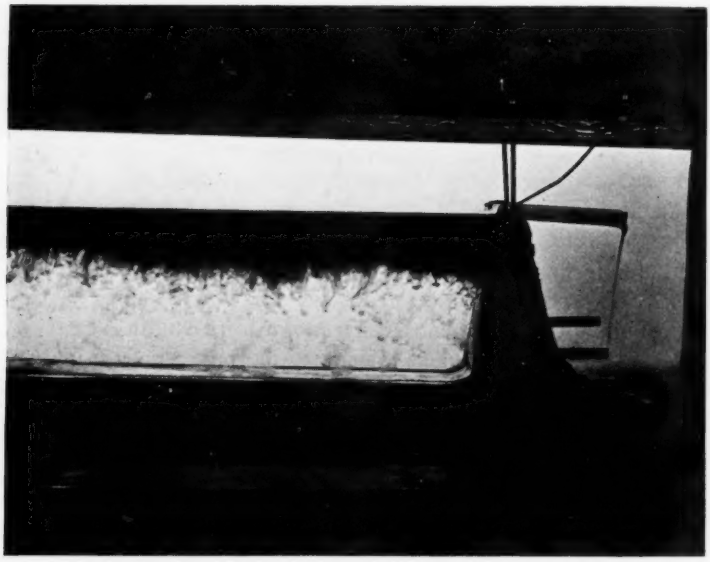


FIG. 12.—Warm muddy water flowing upon the surface of a cold reservoir. As cooling takes place at the interface, small masses of turbid water break through and settle to the bottom of the reservoir, where they continue to flow as a density current.

muddy water is breaking through the interface and settling through the underlying cold water in the experiment illustrated in Figure 12. In the experiment this breaking-through is the result of cooling at the interface; but in the ocean the same effect is produced by wave action, molecular diffusion, and, locally, by bottom irregularities, the propellers of ships, or anything that will cause mixing (Figs. 13 and 14). In the experimental tank the masses of muddy water collect on the reservoir bottom and move downstream as a turbid underflow.

The laboratory technician knows that floccules cause rapid settling of suspended materials, but he also knows that they entrap large quantities of water and that the resulting precipitate behaves as a liquid, flowing readily when the container is tilted, whereas sediment that has recently settled without flocculating refuses to move at much greater angles.[36] The tendency to flow is increased when

FIG. 13.—Small masses of turbid water settling through an underlying layer of salt water. The picture was taken from an airplane flying over the Pacific Ocean near Del Mar, California, February 22, 1941. (Photo by Manly Natland and Roger Revelle.)

coagulation is rapid, as it is almost certain to be when fresh suspensions overflow salty water; and, given an adequate bottom slope, there seems to be a strong possibility that turbid underflows eventually will result when a silt-laden stream overflows the ocean.

At Lake Mead approximately one hundred and forty samples that were analyzed without deflocculation show that, with rare exceptions, from 90 to 98 per cent of the sediment settles with the velocities characteristic of particles from 5 to 20 μ in diameter. Since 77 per cent of the sediment is apparently less than 5 μ in diameter, this

[36] W. C. Krumbein and F. J. Pettijohn, *Manual of Sedimentary Petrography* (New York: D. Appleton-Century Co., 1938) p. 73.

increase in settling velocity might seem adverse for density flows. According to the Stokes law, 5 μ and 20 μ particles settle approximately 0.3 foot and 4 feet per hour, respectively, in still water at 20° C. Such velocities seem almost insignificant when compared with a probable forward movement by the density current of approximately 3,000 feet in the same length of time.

FIG. 14.—A trail of clear water left by a motorboat crossing an area of turbid fresh water on the surface of the Pacific off the coast of southern California. (Photo by Manly Natland.)

THE QUANTITY OF SEDIMENT TRANSPORTED

In the thirteen months prior to the closing of the diversion tunnel at Boulder Dam on May 1, 1936, there were at least four periods of turbid underflow, during which muddy water passed completely through Lake Mead. During the forty-five or forty-six days of turbid discharge an estimated 8,400,000 tons of sediment, at least 90 per cent of which was smaller than 20 μ, flowed from the tunnel. Records of the United States Geological Survey indicate that on a single day during this period the Colorado River carried approximately 1,800,000 tons of sediment finer than 20 μ past the gauging

station at Bright Angel in the Grand Canyon, some 275 miles upstream from Boulder Dam.[37] During September, 1938, it is estimated that the Colorado brought in 23,000,000 tons of similar sediment.[38] That much of this fine material is carried into the lowest parts of the reservoir is shown by the fact that in June, 1941, a submerged lake of muddy water appears to have extended upstream from the dam for more than 35 miles. At that time this lake contained approximately 375,000 acre-feet. Two hundred million tons seems to be a conservative estimate as to the weight of the sediment brought in by turbid underflows—sediment that has been accumulating at the rate of nearly 900,000 tons per week since the closing of the tunnel in 1936.[39]

To estimate the total suspended load transported by density flows produced in natural lakes or in the oceans by inflowing muddy streams, by violent wave action during storms, or by major seismic disturbances would be to indulge in pure speculation, because there seems to be a complete lack of essential data.

Information regarding the transportation accomplished by major dust storms is woefully inadequate but not entirely lacking. Early in May, 1934, such a storm or air mass deposited dust more than halfway across the United States and still contained about 100 tons per cubic mile as it passed over Washington, D.C.[40] In its first few thousand feet above the earth the atmosphere weighs roughly 5,000,000 tons per cubic mile; and, therefore, as the air mass moved over Washington, the dust increased its density only 0.002 per cent. Insignificant though this seems to be, an increase less than 5 times as great has produced density flows in the laboratory.

[37] Grover and Howard, *op. cit.*, pp. 726–28 and 732.

[38] Herbert N. Eaton, "Report of the Interdivisional Committee on Density Currents," presented at the Annual Meeting of the Division of Geology and Geography of the National Research Council, April, 29, 1939.

[39] The estimates given above have been based upon data gathered by the U.S. Geological Survey at its Bright Angel and Willowbeach gauging stations or upon sediment samples secured and measurements made by the U.S. Bureau of Reclamation at Lake Mead. These latter data have been used in conjunction with the area and capacity table for Lake Mead that has been prepared by the U.S. Soil Conservation Service.

[40] I. P. Hand, "The Character and Magnitude of the Dense Dust Clouds Which Passed over Washington, D.C., May 11, 1934," *Monthly Weather Review*, Vol. LXII, No. 5 (1934), pp. 156–57.

At Washington visibility was reduced to half a mile. In 24 of the 192 dust storms which Amarillo, Texas, experienced between January 18, 1933, and February 26, 1936, visibility was reduced to less than 500 feet. Maximum wind velocities reached or exceeded 60 miles per hour in 4 storms, 50 miles per hour in 14, and 40 miles in 82. Storms with wind velocities less than 35 miles per hour were considered to be depositing storms. On the average, they lasted 10 hours, but 7 continued for 50 hours or longer.[41]

In Baca County, Colorado, during 1936 "highway travel was paralyzed and electric-lighted windows were not visible from a distance of 40 feet; while at times the raging 'dusters' cut the visibility to zero, and pedestrians collided with one another in attempting to get about." Aviators reported dust at altitudes of from 15,000 to 20,000 feet during storms that they encountered in Colorado that year.[42]

How much dust is transported by density flows like these? If sediment constitutes only 0.0001 per cent—one millionth—of the volume of such a storm, there will be about 10,000 tons per cubic mile of atmosphere. That is sufficient to increase the density by 0.2 per cent—a very considerable amount for a density current. Possibly the suggested concentration is much too high, and yet a dust storm that struck the central portion of Japan's Hokkaido Island on June 5, 1939, covered 75,000 acres of valuable farm land with an inch of fine silt.[43] Assuming that this unconsolidated deposit had a density of 1.3, it would represent approximately 10,000,000 tons, and this must have been only a fraction of the sediment transported by the storm. Even so, it represents the total load of 1,000 cubic miles of atmosphere in which the dust concentration is 0.0001 per cent by volume.

Undoubtedly, the volcanic debris transported by density currents of various kinds, if expressed in tons, would produce a figure of

[41] Arthur H. Joel, "Soil Conservation Reconnaissance of the Southern Great Plains Wind-Erosion Area," *U.S. Dept. Agric. Tech. Bull. No. 556* (Washington, D.C., January, 1937).

[42] H. F. Choun, "Duststorms in the Southwestern Plains Area," *Monthly Weather Review*, Vol. LXVII, No. 1 (January, 1939), pp. 12–15.

[43] Robert J. Martin, "Duststorms in the Southwestern Plains Area," *Monthly Weather Review*, Vol. LXVII, No. 1 (January, 1939), pp. 12–15.

astronomical proportions. In 1815 the eruption of Tomboro caused darkness for three days at a distance of 300 miles and scattered dust over an area of 6,000,000 square miles. The total volume has been estimated at about 50 cubic miles, or as much sediment as all the rivers of the world carry to the sea in twenty-five years. Even the coarser dust from this eruption and from that of Krakatao in 1883 formed deposits several inches deep at a distance of 1,000 miles. Coseguina, a volcano in Nicaragua, laid down a blanket of dust that was 10 feet thick at a distance of 25 miles and several inches thick at 750 miles.[44] Probably surpassing all of these in magnitude was the eruption of Japan's Asama in 1783—the most frightful eruption on record.[45] Data upon which to base volumetric estimates, however, are not available.

The volcanoes of the Malay Archipelago are credited with the ejection of 300 cubic miles of material between 1770 and 1920, and those in the rest of the world produced an equal amount or perhaps slightly more.[46] Much of this huge total was distributed by density flows, some of which it created, and to all of which it contributed a more or less important part of the driving force.

Certainly the quantity of sediment transported by density currents or available for transportation by them is enormous. It seems equally certain that density flows supply the answer to many problems that the student of sedimentation has found puzzling. For instance, deposits of fine sediments far from shore and at great depths, but containing shallow-water or terrestrial organisms almost exclusively, may not be difficult to explain when consideration is given to density currents.[47] The baffling distribution of ash deposits from certain volcanic eruptions, the thick strata of fine silt or clay laid down far from any source for such materials, the strange scarcity of organic remains of the sort expected on various parts of the ocean floor—these and many other problems are made less perplexing by an understanding of the behavior and the work of density flows.

[44] Blacktin, *op. cit.*, p. 188. [45] Humphreys, *op. cit.*, p. 595.

[46] Amadeus W. Grabau, *Principles of Stratigraphy* (New York: A. G. Seiler, 1924), p. 572.

[47] R. H. Fleming and Roger Revelle, "Physical Processes in the Ocean," *Recent Marine Sediments* (London: Thomas Morley & Co., 1939), pp. 48–141.

CONCLUSIONS

In summing up, it is concluded that—

1. Density currents are among the commonest of natural phenomena.

2. Turbid underflows, dust storms, *nuées ardentes*, and other dust clouds of volcanic origin are gravity flows in which the suspended particles provide at least a part—sometimes all—of the driving force.

3. Density currents, whether atmospheric or aqueous, are important agents in the transportation of sediment.

4. Aqueous density currents in reservoirs are known to carry sediments that contain some particles as coarse as 20 μ. Where the slopes are considerably greater, undoubtedly such currents can carry coarser material. Observations of loess and other wind-borne deposits show that atmospheric density currents move huge amounts of sediments having diameters from 5 to 50 μ.

5. Aqueous density currents transport enormous quantities of fine materials. In American reservoirs the quantity of sediment so transported probably exceeds a million tons per week.

6. Differences in the velocities of atmospheric and aqueous density flows and differences in the conditions under which deposition takes place combine to make atmospheric dust deposits coarser than those laid down by turbid underflows. The great relative coarseness of the sediments transported by density currents that are volcanic in origin is the result of the method of entrainment.

7. The behavior of density flows, in contrast with that of ordinary currents, may be accounted for by the difference in the relative magnitude of the inertia and gravity forces in the two types of flow. In density currents the effective gravity force becomes extremely small; and the inertia forces, since they cannot be diminished, become predominant.

8. By giving due consideration to the mechanism that drives density currents students may be able to clear up some of the baffling problems of sedimentation.

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WIND WORK ACCOMPANYING OR FOLLOWING THE IOWAN GLACIATION

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ABSTRACT

On the Iowan drift wind work is recorded in the etching of the pebbles and boulders, in the development of peculiar ridges known as "paha," and in the deposition of loess that exhibits considerable diversity in texture, color, and degree of ferrugination.

WIND-CUT STONES

The article on "Wind-cut Stones in Kansan Drift of Wisconsin," by Lincoln R. Theismeyer and Ralph E. Digman,[1] prompts me to write concerning the conspicuous development of wind-cut stones on the Iowan drift in northeastern Iowa. They abound on the surface and are involved in the pebbly concentrate that blankets the slopes of that drift. Among these stones was a boulder of white quartzite fully a foot in diameter (Fig. 1) that was clearly shaped into dreikanter form.

The effect of wind action in this region was noted by W. C. Alden and M. M. Leighton in their report on the Iowan drift,[2] but this action was interpreted by them to have been of brief duration and owing to some extent to the anticyclonic winds coming from the ice sheet. The exceptional thickness of the loess near the border of the Iowan drift was also attributed by them to such winds. In my opinion the shaping of such a boulder as that shown in Figure 1 may require a more protracted period of wind action than that covered by anticyclonic winds. The amount of pebbly concentrate also seems greater than is likely to have been produced in connection with such winds. Judging from the proportion of coarse pebbles and boulders

[1] *Jour. Geol.*, Vol. L (1942), pp. 174–88.

[2] "The Iowan Drift," *Iowa Geol. Surv.*, Vol. XXVI (1915), pp. 50–212, esp. p. 156; see also G. F. Kay, "Origin of the Pebble Band on Iowan Till," *Jour. Geol.*, Vol. XXXIX (1931), pp. 377–80, and W. H. Hobbs, "Loess, Pebble Bands, and Boulders from Glacial Outwash of the Greenland Continental Glacier," *Jour. Geol.*, Vol. XXXIX (1931), pp. 381–85.

in the subjacent till, it represents the removal of several feet of till. It seems to be the product of long-continued slope wash. However, the relative effectiveness of wind work and slope wash in this district remains to be determined.

As to the article by Theismeyer and Digman on ventifacts in the Arnott moraine, I am sorry to have to make the comment that the

FIG. 1.—Boulder of white quartzite, with dreikanter shaping by wind, on the surface of the Iowan drift in southwest part of Dubuque County, Iowa. (Photo by Samuel Calvin, taken in 1907.)

Arnott moraine in all probability is of Illinoian age, as is also the Marshfield moraine, both being classed by Weidman as "Second drift." The amount of erosion, as determined by a special study of the Marshfield moraine that I made in 1923, is scarcely one-third as great as the general amount of erosion of the Kansan drift. It shows a degree of erosion similar to that of the Illinoian drift in Illinois.[3]

In his brief article in this *Journal*, published in 1931,[4] Hobbs

[3] See "Quarternary Geology of Minnesota and Parts of Adjacent States," *U.S. Geol. Surv. Prof. Paper 161* (1932), pp. 22–23.

[4] *Op. cit.*

called attention to the powerful sand-blast action on boulders in the glacial outwash bordering the Greenland ice sheet. The etched boulders there bear a striking resemblance to etched erratic boulders that are a common feature on the surface of the Pleistocene drift sheets. As a result of discussing this matter with him in connection with the preparation of this note, he has kindly written a fuller statement about the distribution of such boulders and has explained more fully the wind work on the border of the ice sheet.

PAHA RIDGES

The Iowan drift sheet is the type area of the paha brought to notice by W. J. McGee in his "Pleistocene History of Northeastern Iowa."[5] They are also discussed in more detail by members of the Iowa Geological Survey in their county reports. These canoe-shaped ridges are in large part of wind-deposited material, much of it a loose-textured loess. They have a remarkably uniform trend, about W.NW.–E.SE., throughout their occurrence from near Cedar Rapids, Iowa, eastward into the edge of Illinois, some of the best ones being in Whiteside County, Illinois, as shown in Figure 2. They seem to call for a pronounced dominance of eastward-moving winds for a considerable period.

As these ridges also conform pretty closely in trend with the general direction of movement of the Iowan ice sheet in that district, this naturally raises the question whether the anticyclonic winds from the ice sheet may have played an important part in forming them. Their extension into Illinois also raises the question of crossing the site of the Mississippi River by the Iowan ice sheet. The Iowa Geological Survey has found the Iowan drift in a strip 12–15 miles in width that extends clear to the Mississippi River in the section between Clinton and the valley of Wapsipinnicon River. This is directly opposite the middle part of the paha district in Illinois. My studies were completed in 1897, about thirteen years before the anticyclonic winds from the continental ice sheets had been recognized and definitely brought to notice by Professor Hobbs.[6] I was,

[5] *U.S. Geol. Surv. 11th Ann. Rept.*, Part I (1891), pp. 199–577.

[6] "Characteristics of the Inland-Ice of the Arctic Regions," *Proc. Amer. Phil. Soc.*, Vol. XLIX (1910), pp. 57–129; see also W. H. Hobbs, "The Pleistocene Glaciation of North America Viewed in the Light of Our Knowledge of Existing Continental Glaciers," *Amer. Geog. Soc. Bull. 43* (1911), pp. 641–59.

however, of the opinion that the Iowan ice sheet had some sort of controlling influence in the development and extent of the paha ridges. For that reason I mapped them as "Ridged loess and sand

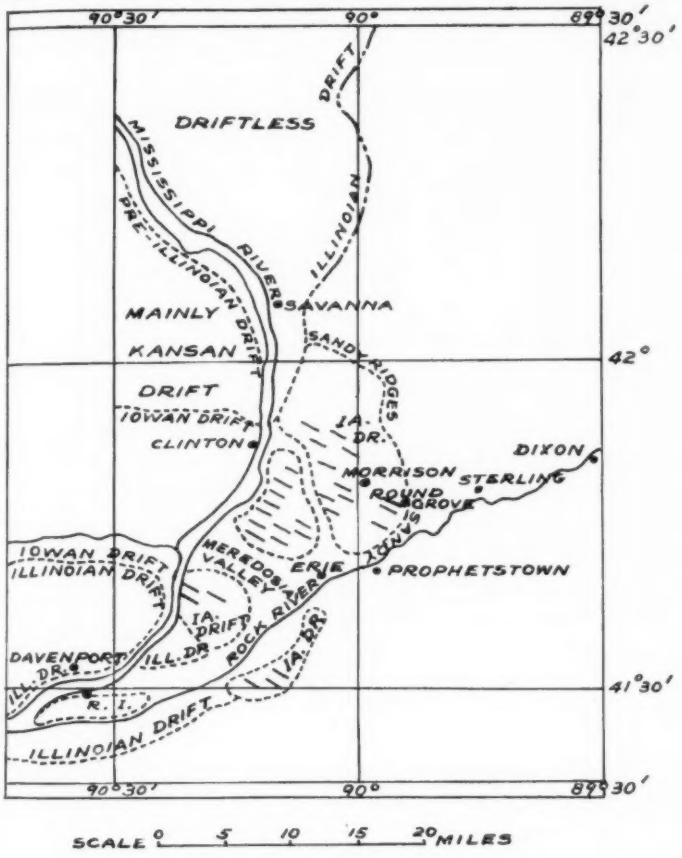

FIG. 2.—The Paha district in Illinois

on the Iowan drift border" and as "Sand and loess ridges on or near the Iowan drift border" on Plate XII of my report on the Illinois Glacial Lobe. I am now presenting that part of my map as Figure 2, the only change made being the extension of the area about 6 miles northeast of Morrison to take in a ridging that appears on the Mor-

rison topographic map. I now think we may safely infer that the anticyclonic winds have had a strong influence in their development and extent.

The Iowan drift is a thin and patchy deposit for some distance west from the Mississippi River, and so it is natural to expect that it

FIG. 3.—Photo by Samuel Calvin in 1907. Shows the contact (*C–C*) between highly ferruginated blue loess with the dark spots and overlying brown loess of loose texture. Location in Section 6, Whitewater Township, Dubuque County, Iowa, at east end of the "Worthington Lobe" of Iowan drift.

may be still more scanty east of the river. There is a possible Iowan patch about 10 miles east of Clinton on upland fully 200 feet above the Mississippi River, as shown by contours on the Morrison topographic map. The leaching is scarcely one-third as deep as on the adjacent Illinoian, and the iron staining is much fainter. West and south of Morrison, on the borders of Rock Creek and eastward from its valley, sandy till fresher than the Illinoian was noted at several points. In some cases it was found to be calcareous to within 3 feet

of the surface. A short distance west of Round Grove, excavations examined in 1897 showed a soft till, yellow at top, but changing to bluish color within 3 or 4 feet of the surface. There also is a fresh till exposed south of Rock River, where the wagon road ascends toward Spring Hill; in this exposure the loess rests directly on an unleached

FIG. 4.—Laminated loess exposed on west line of Section 23, Dodge Township, Dubuque County, Iowa, between the Worthington and Farley tongues of the Iowan drift near their place of separation. (Photo by Samuel Calvin in 1907.)

till surface. These occurrences of a till fresh enough to compare closely with the Iowan in Iowa lead me to think that the Iowan ice sheet may have reached nearly as far as the paha ridges. Exposures of till, unfortunately, are rare in this district partly because of the shallowness of the valleys and partly because of concealment by loess and sand. There are sandy ridges that may be outside the Iowan limits on elevated land southeast of Savanna, whose position is indicated in Figure 2. On their southwest side the loess is scanty compared with that to the north and east, but most of the till seems to be of Illinoian age.

THE LOESS

The loess exhibits considerable diversity in texture, color, and degree of leaching and ferrugination in certain areas, notably in western Dubuque County, Iowa, where it borders loess-free areas that were interpreted by Calvin and Bain to have been occupied by the Iowan

FIG. 5.—Ferruginous band separating blue loess from overlying loose-textured brown loess. The material in the band suggests derivation from the blue loess by an upward movement of moisture. Location in Section 14, Taylor Township, Dubuque County, Iowa, at the east end of the Farley lobe of Iowan drift, a mile south of Epworth. (Photo by Samuel Calvin in 1907.)

ice tongues. In several places a blue loess, full of ferruginous nodules and tubes and of compact indurated character, underlies a loose-textured brown loess, as shown in Figure 3. The blue loess evidently was a poorly drained deposit, apparently for a considerable time a surface deposit carrying a growth of vegetation, the roots of which penetrated the deposit. After decay, the resulting cavities were filled with the ferruginous material. The overlying loose-textured brown loess is not a poorly drained deposit and presumably is of

markedly later date of deposition. In Figure 4 the upper loess shows distinct lamination. The texture is that of a silt loam rather than a fine sand. It occurs where the Worthington and Farley tongues of Iowan drift are represented by Calvin and Bain to become distinct tongues, in section 23, Dodge Township, Dubuque County. It presumably is a markedly later deposit than the underlying blue loess.

The photo showing a ferruginous band separating the blue loess from the overlying brown loess (Fig. 5) was taken at the east end of the Farley tongue, south of Epworth. It is of interest to determine whether this band was derived from the blue loess by an *upward* movement of moisture from that deposit. The material in the band seems to favor that derivation.

WIND—THE DOMINANT TRANSPORTATION AGENT WITHIN EXTRAMARGINAL ZONES TO CONTINENTAL GLACIERS

WILLIAM HERBERT HOBBS
University of Michigan

ABSTRACT

The dominant transportation agent within the extramarginal zones to continental glaciers is for most of the time exclusively the fierce anticyclonic storm wind directed outward normal to the glacier front. Transportation by running water during the remainder of the year is almost exclusively by meltwater. The volume of this water over the outwash plain fluctuates between trickling braided streams and floods covering the plain on warm days, so that the boulders carried in small icebergs from the glacier front either are buried within the deposits or are left stranded upon the plain. During the longer period of wind dominance there result pebble bands (pavements), ventifacts, etched boulders, dunes, and loess deposits—all of them extramarginal to the glacier.

On the basis of studies made near the border of the Greenland continental glacier I called attention in this *Journal* for May-June, 1931, to the erroneous conception which has long prevailed regarding the conditions which have obtained outside the Pleistocene continental glaciers of North America and Europe.[1] At the Cambridge meeting of the American Association for the Advancement of Science in 1933, I again referred to the Greenland observations and discussed the evidence that the same conditions had been present about the North American continental glaciers of Pleistocene times. The traditional view, based upon studies of mountain glaciers like those of the Alps, had taken no account of the fierce storms which blow outward off the ice of continental glaciers or of the restriction of drainage to a brief warmer season during which meltwater issues from beneath the glacier. Meteoric precipitation is almost exclusively of snow or rime.

As soon as the brief warm season has come to an end, the agent of extraglacial transportation is no longer running water but strong wind currents directed radially outward from the glacier, or nearly

[1] "Loess, Pebble Bands, and Boulders from Glacial Outwash of the Greenland Continental Glacier," *Jour. Geol.*, Vol. XXXIX (1931), pp. 381–85; cf. also "The Origin of the Loess Associated with Continental Glaciation Based on Studies in Greenland," *Compt. rend. Congrès internat. géog.*, Vol. II (Paris, 1931), pp. 1–4.

at right angles to the glacier front. During the brief warm season the glacier front becomes more or less undermined by the thaw water which has descended within the marginal crevasses of the ice and issues from beneath the front. Masses of the lower ice layers, loaded with englacial rock debris, including large boulders, become detached as small bergs which are not too heavily burdened to float off in the braided streams on the outwash plain. During exceptionally warm days the streams of outwash may coalesce and become transformed into lakes which may cover most of the plain. When the temperature drops, the flood subsides, and bergs with their enclosed boulders are stranded over the plain away from the braided channels.

When the warmer season ends, no meltwater issues from the glacier, and the entire plain of outwash quickly dries. This marks the beginning of the longer winter season, and wind takes over from the thaw water the work of transportation.

These conditions within the extramarginal zones of the continental glaciers are at this time similar to those within the deserts far removed from glaciers. The rock materials of the outwash, varying in degree of coarseness—rock flour, sand, pebbles, and boulders—are sorted out by the winds. Dust, sand, and the smaller pebbles are lifted and carried away, leaving behind an armor of pebble pavement to protect the materials below, and in every way this pavement is similar to the *sêrir* of low-latitude deserts. The boulders stranded on the plains are now exposed to the driving sands and in consequence become deeply etched, while the larger pebbles and even boulders of uniform hardness become smoothed, planed, and transformed into ventifacts. Those boulders which are crossed by dikes or veins of harder material under this drilling action develop into the "stone lattice" familiar from the low-latitude deserts, and here are found the finest examples anywhere known. Igneous rocks of porphyritic texture show the phenocrysts in strong relief. Conglomerates containing quartzite pebbles show the pebbles projecting from the general surface, whereas pure quartzite rocks may be transformed into smooth ventifacts.

The sand, after accomplishing its work of planing and etching, comes to rest in dunes which under the fierce winds are ridges normal

to the glacier front, while the dust is widely spread as loess deposits, sometimes with pebble bands at their base and also in some cases with included etched boulders.

All these types—pebble pavements, sand dunes, loess deposits, deeply etched boulders, pebble bands, and ventifacts—are now known to be widely distributed throughout the glaciated area of North America, but until recently they have found little mention in the literature of the subject. At the time my first paper on the subject was issued (1931), I sent out widely to American geologists and

FIG. 1.—Deeply etched boulders from the glaciated area around Ann Arbor, Michigan

others throughout the glaciated area of North America "A Call for Information concerning Etched Erratic Boulders."[2] This call brought a sufficient number of responses to show that similar etched boulders are common within many parts of the glaciated area of North America, particularly likely to be found selected as curiosities in house yards (see Fig. 1) or even as gravestones in cemeteries. I have myself found such etched boulders by the score within the glaciated area surrounding Ann Arbor, Michigan, and during subsequent travel I have seen them at many other localities.

The restriction of American and European loess deposits to extra-

[2] *Jour. Geol.*, Vol. XLIII (1935), pp. 551–52. The inquiry has not been concluded and further communications are requested for the completion of a map of distribution for the glaciated area of North America which is now in preparation.

marginal zones about Pleistocene continental glaciers had long been recognized, and Dr. George F. Kay has treated of sand dunes, pebble bands, and etched boulders in the loess on Iowan till.[3]

There is need for revision of studies of much of the glaciated areas of North America with this altered picture of extramarginal conditions around the glaciers in mind.

[3] "Origin of the Pebble Band on Iowan Till," *Jour. Geol.*, Vol. XXXIX (1931), pp. 377–80; see also Lincoln R. Thiesmeyer and Ralph E. Digman, "Wind-cut Stones in Kansan Drift of Wisconsin," *Jour. Geol.*, Vol. L (1942), pp. 174–88.

REVIEWS

A Catalogue of the Foraminifera. By BROOKS F. ELLIS and ANGELINA R. MESSINA. New York: American Museum of Natural History Special Publication. Pp. 30,000; illustrated. $100.

Scientists have been writing about the Foraminifera for at least two centuries—an even longer period than the authors of the *Catalogue* conservatively state in their Preface. Papers have appeared in many different publications, in many lands, and in diverse languages; moreover, they have included all sorts of systems in naming, describing, and illustrating the various species. To make matters worse, to this very day some perverse writers fail to follow completely the International Rules of Zoölogical Nomenclature. As a result, the literature on the Foraminifera has become not only amazingly extensive but astonishingly confused. Inasmuch as the Foraminifera loom larger and larger as an important group not only with scientific but with utilitarian significance, it has been recognized for some years that the preparation of a general catalogue would tremendously facilitate work with these protozoans. The enormous size of such an undertaking, however, prohibited any one person or any single institution from subsidizing the effort.

In 1927 Dr. Ellis began to worry about this problem, and he concluded that the facilities to compile such a catalogue could best be found in the libraries in and adjacent to New York City. In 1930 Dr. Ellis joined the faculty of the department of geology of New York University. Thus the requisite library facilities became available, and he boldly started on the catalogue as an individual project. Shortly thereafter Miss Messina, then a student assistant, joined the undertaking, and by 1932 the broad framework of the *Catalogue* began to stand out. By this time both New York University and the American Museum of Natural History had become sufficiently interested to further the work, but funds were still not in sight to complete the project. In 1933 the Gibson Committee was organized to give temporary employment to men and women out of work in New York, and an agreement was reached to provide a number of clerical workers for the foraminiferal catalogue. When the federal government took over this relief work, an application was made for the support of a larger project. This application also was approved, and by early

1934 a staff of eighty-two workers had been assembled. New York University provided the space, the equipment, and the supplies, and the American Museum of Natural History either provided or obtained the literature necessary for the undertaking. By 1935 the project had been included in the Works Progress Administration, and the personnel was increased to the amazing total of 175. Mayor La Guardia became the official sponsor of the project, and the two institutions previously mentioned were named as cosponsors.

In June, 1940, the trustees of the American Museum voted to establish a department of micropaleontology. Thus at long last Dr. Ellis' dream became a reality. Not only were the first volumes of the *Catalogue* issued, but a department responsible for its continuance had been provided for. Undoubtedly, there thus culminated one of the largest nonindustrial research projects ever completed under a unified guiding group. The results not only constitute a crowning achievement for the authors but for all of the many persons who worked on the project and for the various organizations that facilitated its successful completion. In spite of the size of the army of workers involved in this research, Charles B. Lawrence, Jr., the manager of the planning department of the W.P.A., was probably not too inaccurate when he stated, in the Preface, "The *Catalogue of Foraminifera* . . . . was compiled with an expenditure of professional hours probably totalling less than those that will be saved for science during the first few months of its use."

The purpose of the *Catalogue* is to present in uniform arrangement all the essential data for each genus, species, and variety of Foraminifera. The policy of the authors has been to present facts rather than opinion, and they have thus commented only where such comment seemed essential. All species considered to be new by their respective authors have been listed, regardless of the fact that many such "species" names have been suppressed as synonyms. Obvious homonyms have also been included along with all the new names, but non-Linnean designations and genera and species described in languages with non-Latin alphabets have been excluded. Each genus, species, or variety listed in the *Catalogue* is considered a unit, and they are arranged alphabetically by genera and alphabetically by species within each genus. The Index is made up in the same fashion and is cross-referenced to show synonymy according to authors. Moreover, in order to permit rearrangement and to provide for the insertion of species to be erected in the future, etc., the *Catalogue* is bound in ledger-type post binders. All figures originally used to illustrate each

species or variety are reproduced with the original figure numbers. Under each description of a type the original phraseology of the author is quoted.

Although the *Catalogue* is a complete unit within itself, the American Museum has also established continuing memberships at $25 per year. Continuing members are entitled to receive the new material as issued and have access to all the source material in the archives. In addition, extensive researches will be made by the staff on request, and the charge for this sort of work will be nominal. Photostats and photographs also will be supplied to members at standard prices.

The volumes of the *Catalogue* are well bound and attractively printed. It seems to the reviewer, however, that some of the reproductions leave considerable to be desired, despite the fact that it is recognized that the originals from which they are taken in many instances provided poor material with which to work. The fact that the thirty volumes will inevitably have their contents slightly rearranged in position as a result of the insertion of new material has made it impossible to indicate on the back of each the first and last genus included. Owners of sets can obviate this minor but irritating difficulty, however, by pasting a readily changed finding key on the back of each volume.

The word "monumental" is doubtless very much overworked by reviewers, but if the adjective can ever be used appropriately it is certainly apropos as applied to this particular *Catalogue*. The thirty volumes, with their thirty thousand odd pages, require several yards of shelf space. There is no doubt, however, that all who have been fortunate enough to acquire this invaluable set are more than pleased to give it room, and that many others for whom the price at present seems prohibitive are hopefully saving shelf space in the sanguine expectation that a windfall in the way of the necessary library appropriation will be forthcoming.

CAREY CRONEIS